Volume 1, Number 1 February 1993

ISSN: 0869-5938

STRATIGRAPHY AND GEOLOGICAL CORRELATION

Official English Translation of *Stratigrafiya, Geologicheskaya Korrelyatsiya*

Editor-in-Chief
Boris S. Sokolov

Academician and Councellor of the Presidium
Russian Academy of Sciences

Journal of the Fundamental and Applied Aspects of Stratigraphy and the Correlation of Geologic Events and Processes in Time and Space

INTERPERIODICA PUBLISHING

Stratigraphy and Geological Correlation

(*Stratigrafiya. Geologicheskaya Korrelyatsiya*)

Published 6 times a year
by Interperiodica (Russia) and the Russian Academy of Sciences
Distributed worldwide by Interperiodica
ISSN: 0869-5938

Editor-in-Chief
Boris S. Sokolov
Counsellor of the Presidium of Russian Academy of Sciences,

Address for correspondence: Staromonetnyi Pereulok, 22, Moscow, 109180, Russia
Phones: (7 - 095) 231 - 2164. Fax: (7 - 095) 233 - 5590.

Deputy Editor-in-Chief
Mikhail A. Semikhatov
Corresponding Member, Russian Academy of Science,
Laboratory Chief, Geological Institute,
Russian Academy of Science

Assistant Editor
Aleksei B. Herman
Senior Scientist,
Geological Institute,
Russian Academy of Science

EDITORIAL BOARD

Mikhail A. Akhmet'ev
D. Sc. (Geol.-Min.),
Laboratory Chief, Geological Institute,
Russian Academy of Sciences

Aleksandr S. Alekseev
Professor, Chair of Paleontology,
Moscow State University

Mikhail N. Alekseev
D. Sc. (Geol.-Min.),
Chief Scientist, Geological Institute,
Russian Academy of Sciences

Ivan A. Basov
D. Sc. (Geol.-Min.),
Deputy Director, Institute of Lithosphere,
Russian Academy of Sciences

Michael G. Basset
Professor,
National Museum of Wales, Cardiff, UK

William A. Berrgen
Professor,
Department of Geology and Geophysics,
Woods Hole Ocean Institute, USA

Elena V. Bibikova
D. Sc. (Geol.-Min.),
Chief Scientist,
Vernadskii Institute of Geochemistry
and Analytical Chemistry
Russian Academy of Sciences

N A. Bogdanov
Corresponding Member,
Russian Academy of Sciences,
Director, Institute of Lithosphere,
Russian Academy of Sciences

Nikolai M. Chumakov
D. Sc. (Geol.-Min.),
Chief Scientist, Geological Institute,
Russian Academy of Sciences

Yurii B. Gladenkov
D. Sc. (Geol.-Min.),
Deputy Director, Geological Institute,
Russian Academy of Sciences

Jurgen Remane
Professor,
institute de Géologie,
Université de Neuchatel, Switzerland

Dmitrii L. Kaljo
Professor,
Institute of Geology, Tallinn, Estonia

Victor E. Khain
Academician,
Russian Academy of Sciences,
Chief Scientist, Institute of Lithosphere,
Russian Academy of Sciences

Kiyotaka Chinzei
Professor,
Geological Institute,
University of Tokyo, Tokyo, Japan

Lidiya A. Nevesskaya
D. Sc. (Biol.),
Laboratory Chief, Paleontological Institute,
Russian Academy of Sciences

Aleksandr G. Ponomarenko
D. Sc. (Biol.),
Chief Scientist, Paleontological Institute,
Russian Academy of Sciences

Aleksei Yu. Rozanov
D. Sc. (Geol.-Min.),
Director, Paleontological Institute,
Russian Academy of Sciences

Boris A. Sokolov
D. Sc. (Geol.-Min.),
Professor,
Dean of Geological Faculty,
Moscow State University

Otto Walliser
Professor,
Göttingen University, Göttingen, Germany

Sun Weigo
Professor,
Nanjing Institute of Geology
and Paleontology, Academy Sinica, China

Victor A. Zakharov
D. Sc. (Geol.-Min.),
Laboratory Chief, Institute of Geology,
Geophysics, and Minerology,
Siberian Branch, Russian Academy
of Sciences, Novosibirsk

Aleksandr I. Zhamoida
Corresponding Member,
Russian Academy of Sciences,
Russian Geological Institute,
St.-Petersburg

Staff Editor Tamara V. Trishkina

SCOPE

Stratigraphy and Geological Correlation (Stratigrafiya. Geologicheskaya Korrelyatsiya) is the only journal in the world designed to provide comprehensive coverage of the fundamental and applied aspects of stratigraphy and the correlation of geological events and processes in time and space. Articles are based on the results of multidisciplinary research, and are directed toward researchers, university professors, students, and geologists interested in understanding stratigraphy and the chronological features of the world's geological record.

SUBSCRIPTION OFFICE
INTERPERIODICA
P.O. BOX 1831, BIRMINGHAM, AL 35201-1831
PHONE: (800) 633-4931; (205) 995-1567 (outside US and Canada); FAX: (205) 995-1588

SUBSCRIPTION PRICE
US $540 (North America), $576 (other). A 5% discount is given on 2-year orders. Subscriptions may be entered at any time. Orders may be placed at agencies or directly. 7.5% discount to all subscription agencies. Cancellations not allowed except for duplicate payment. Prepayment required. Advance payment required in US$/US bank. MasterCard/Visa accepted. Checks payable to the name of the journal or Interperiodica.

The journal was founded in 1993.
Original Russian Edition Copyright © 1993 by *Stratigraphy and Geological Correlation.*
English Translation Copyright © 1993 by Interperiodica.

THIS PUBLICATION IS PRINTED ON ACID-FREE PAPER

CONTENTS

Volume 40, Number 1, 1993

To the Readers of the Journal Stratigraphy and Geological Correlation
Boris S. Sokolov, Editor-in-Chief 1

General Subdivision of the Precambrian: Comparison of the Latest Scales
M. A. Semikhatov 4

Riphean Middle Siberian Glaciohorizon
N. M. Chumakov 17

The Age Boundaries of the Early Precambrian Geological Evolution of the Enisei Ridge
E. V. Bibikova, T. V. Gracheva, V. A. Makarov, A. D. Nozhkin 29

Microbiotas of the Kerpyl' Group in the Riphean Hypostratotype of Siberia
A. F. Veis and N. G. Vorob'eva 35

Main Events in Permian History of the Tethys and Fusulinids
E. Ya. Leven 51

Bashkirian Stage of the Sim Brachysyncline (the Western Slope of the South Urals)
R. M. Ivanova and B. I. Chuvashov 66

Stages and Cycles in the Late Cretaceous Floral Changes of the Anadyr'-Koryak Subregion (Northeast Russia)and Their Connection with Climatic Changes
A. B. Herman 77

Paleogene Geological Events in Kamchatka
Yu. B. Gladenkov and A. E. Shantser 88

Zonal Subdivision of the Upper Toarcian in the Northeast of Russia
Yu. S. Repin and I. V. Polubotko 99

Correlation of the Miocene Mammalian Reference Localities with the Magnetochronologic Scale
E. A. Vangengeim and M. A. Pevzner 108

Lower Oligocene Palynology and Phytoplankton of the Northern Caucasus
N. I. Zaporozhets 117

The Pontian-Caspian Early Pleistocene
P. V. Fedorov 125

Stratigraphy and Geological Correlation, Vol. 1, No. 1, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 20, No. 2, 1993, pp. 3 - 5.
Original Russian Edition Copyright © 1993 by Stratigrafiya. Geologicheskaya Korrelyatsiya, Sokolov.
English Translation Copyright © 1993 by Interperiodica.

To the Readers of the Journal *Stratigraphy and Geological Correlation*

The idea of establishing an academic journal under this or similar name started long ago. There were at least two reasons behind it. In the first place, in 1956, geological journals started to become specialized. Until then, only one geological journal, *Izvestiya AN SSSR, Seriya Geologicheskaya* by the USSR Academy of Sciences, existed since 1936. Such separation reflected the differentiation of geological sciences, and the interests of geologists of that time, including stratigraphers. In the second place, stratigraphy, a base field of different geological studies, was expanding its methodology. Improvements in the principle of regional and global correlation, as well as more universal approaches to the general goals and tasks of stratigraphy, were necessary. It became vital to join international efforts to determine such common stratigraphic standards as the base for tipization of boundaries of all ranks, ranges of subdivisions and their hierarchy, nomenclature, criteria of correlation control, etc. As a result, the number of stratigraphic papers drastically increased, as did the diversity of geochronological information.

The success in spatial stratigraphy and isotope geochronology shed new light on the phenomenon of geological time and practical implementation of its material carriers. An idea was developed in the late 1960s to establish a single research program for a historical geological analysis which could unite various globally distant geological phenomena, events and processes, so that their space-time relations could be established accurately and reliably.

The idea was the essence of geology, its link with history. This was the way in which the principal statements of the International Geological Correlation Program (IGCP) were formulated. This program is the most important one for decades. Today, it embraces a great number of global and large-scale regional projects involving all continents and the World Ocean. The projects raise modern geology to a new level. In this country, the program has become very prominent since its inception.

Opening this new journal, we hope that problems of stratigraphy and geochronology will find a proper place on its pages (a journal of this kind has never existed here so far). We also hope that discussions of these problems will be based on the general concepts of the IGCP, and primarily on the correlation of geological and biological events in the lithosphere and biosphere of the planet, which reveals regularities in the evolution of geospheres in general. In this respect, the well-known statement from the *Closing Report of the IGCP Preparation* (Paris: UNESCO 1972) is significant. It says: "Stratigraphic classification and correlation is the basis of the synthesis of many disciplines."

The journal *Stratigraphy and Geological Correlation* was initiated in accordance with the decision of the Russian Academy of Sciences (RAS) on April 21, 1992. It was established as a journal of the Presidium of RAS and its Department of Geology, Geophysics, Geochemistry and Mining Sciences. Issues will be above 100 pages, and will be published six times a year. The decision specified that the journal "will publish results of all historical geological studies on a multidiscipline basis, because of the achievements gained by stratigraphy and time-space correlation of geological events and processes. The journal will include articles on general, theoretical and regional problems of stratigraphy, involving continents, sedimentary basins, shelves, and the sedimentary cover of the World Ocean. It will publish papers on methods of stratigraphic research (bio-, eco-, litho-, magneto-, chemo-, seismo-, and holostratigraphy as a whole). There will be mate-

rials on geochronology and isotope geochronology; history and evolution of the biosphere, including abiogenic factors; aspects of geological correlation; and global geohistorical changes in the Earth."

A journal of such scope will interest a wide circle of experts and scholars investigating diverse documentation and evidence about the geological past of our planet. Complete historical geological reconstruction is needed to comprehend the evolutionary trends and crises of the past. Such knowledge is very important to scientifically substantiate development of lithosphere mineral resources.

The profile of the journal has been proven to attract foreign readers as well. Therefore, it will be simultaneously published in two versions: in Russian by the *Nauka Publishers* of the RAS, and in English by the joint Russian-American publishing house *Interperiodica*. The editorial board of the journal will also be international.

It is easy to see that there is a high expectation from the research on the Earth's sedimentary cover, its genetic and regional types, different climatic zones, diverse paleogeographic environments, and geodynamic regimes of geological evolution. It will be significant to consider the entire time range of such events, from the oldest units of aquatic sedimentary and biochemogenic origin, to the latest Holocene cover of continents and of the ocean floor. The history of the stratisphere spans almost four billion years. It holds unique information for fundamental studies in stratigraphy, its hierarchic geochronological system with its own material expressions (subdivisions, boundaries). Such information helps in understanding the nature of geological time-and-space, periods and rhythms of the biogeospheric process, and how it is impacted by events in outer space. Lastly, it is vital for comprehending the evolution of the biosphere, and all its biotic and abiotic parameters.

Today, one can reliably assert that the Earth is the only cosmic object in the Solar system which contains a complete record of events for several billions of years, because its unique structure developed such a "condenser-chronograph" as its sedimentary cover, or stratisphere, which is mosaic but, nevertheless, chronologically complete as a whole, and is the most informative part of the Earth's crust. This information can be systematized in many ways, though its scientific value is primarily assessed by spatial correlation and by revealing time boundaries of changes of various ranks. Any new data and results of their generalization on this subject will be of paramount interest for our journal.

Most of the knowledge in historical geology was accumulated during the two centuries after the paleontological method was first introduced, when more or less complete fragments of the stratispherical cover of the continents and their intercontinental correlation were investigated. The most complete chronological history of the Earth is the history of its continents from the ancient Archean, when its biosphere was still anaerobic and rudimentary, to the Late Quaternary.

Modern concepts of lithospherical dynamics suggest that the intercontinental correlation is not only sufficiently complete chronologically, but is also of great value. In any case, we can hardly hope to discover under the vast ocean, beyond the continental shelves, any unknown links of Precambrian or Paleozoic complexes, which connected the segments of the sedimentary cover, but have now disappeared. At the present level of biostratigraphy and of stratigraphic correlation, it is possible to assert, for example, that the best-studied Paleozoic complexes of contemporary continents have no chronostratigraphic lacunas which could be filled from the lithosphere beneath the World Ocean.

However, our ideas on the complete makeup of the Mesozoic and Cenozoic cover of continents radically changed in the late 1960s, after the first deep-sea drilling operations made some fundamental discoveries. It was found that the Mesozoic and Cenozoic cover also embraces the ocean floor. However, its base, resting right on the basalt basement, regularly contains deposits which are usually not older than the middle Mesozoic.

The preceding boundary event was significant for geological history. It was only after that event that the Earth could have conditions for such widely spread transgressive covers and for post-Hercynian sedimentation basins of a new type. Because of these discoveries and generalizations, geologists could for the first time make their correlation at a genuinely global scale. Such an approach, based on events and their correlation, to comparative geological analysis of the World Ocean sedimentary cover, of the zones transitional to continents, and of the sedimentary cover of continents, is a trend of research started only recently. It has already given birth to new concepts in stratigraphy and methods of correlation. It promises to provide us with important information for the theory and practice of stratigraphy, especially regarding the economic development of shelf seas.

Another field of study which expectedly will be discussed in the journal, deals with old sedimentation basins of various paleogeographic types, biotic environments, and tectonic regimes in their complete evolutionary integrity. Historically, the fundamentals of regional stratigraphy were based upon investigations of the above. Regional stratigraphy became a starting point for global stratigraphic issues, and the standard for international stratigraphic and geochronological scales. After the IGCP was adopted, it took on the responsibility of improving this as the supporting framework of correlation. Despite great progress and the well-coordinated activities of international groups investigating geological systems, there is much room for improvement.

As in the past, regional basins and their comparison remain in the focus of all studies. Productive oil- and gas-bearing basis are particularly important. The high-

est possible accuracy in the most detailed subdivision of oil- and gas-bearing deposits, as well as their stratigraphic correlation, has been the basins for new ideas and numerous methods in stratigraphy. Researchers now expect to get the most important results from the complex "basin stratigraphy," which regards the entire ecological facies structure of basins, the physical parameters of their environment, and conditions of their evolution.

It is not only the stratisphere on the Earth that is unique, but also its biosphere, to which the former greatly owes for its origin. The stratisphere and biosphere of the geological past (the "bygone biospheres," according to V.I. Vernadskii), are sources of mutually supplemental information. They are the basis for comprehending and reconstructing the environment and life systems, representing hundreds of millions and billions of years of geological history. This never ending component of the biospheric process helps in understanding biospheric evolution and its characteristics, the crises and rearrangements in ecosystems, interrelations between the changing earth biota and the alternating conditions in the atmosphere, hydrosphere, and the landscape. Investigation of these processes by itself is very important. But today their significance has increased in connection with geological prediction, because they allow us to see how biota of the past survived critical situations. Due to these investigations, we can assess the role played in the evolution of ecosystems by such factors as geological duration, climatic changes, changes in the composition of the atmosphere and waters, tectonic activity, changes in the composition of biota itself, namely, in its vegetation, animal, bacterial components (perhaps, the latter component should be regarded as the first one). Knowledge, not hypotheses, has become paramount because of the ever-growing necessity to take protective measures against the anthropogenic impacts upon the biosphere. It is also vital to understand the natural stability of the biosphere as a most important ecosystem.

The condition of ecology and the fate of the biosphere have become a world-wide concern among scientists. They feel it is vitally important to advance current global programs studying the natural environment and climate. The Russian Academy of Sciences has already initiated multi-faceted investigations of this kind. Among these, the most relevant to the journal are the studies of rearrangements in the Earth's ecosystems and the evolution of its biosphere, including all its biotic and abiotic factors. The reliability of historical control here is provided by stratigraphy and geological correlation.

Although having the word *stratigraphy* at the beginning of the Journal's title, the Editorial Board certainly did not intend to restrict the Journal to stratigraphic problems. It was primarily their tribute to the logics of space-time correlation of all geological processes and events (including tectonic, sedimentological, paleobiological, etc.), which occurred on the Earth in the geological past. Any global or regional event corresponding to the general concept of the Journal will find a proper place in it.

This brief introduction to the journal *Stratigraphy and Geological Correlation* has only touched upon some of the most important trends in research. However, it indicates the broad scope of historical geological investigations and information which the Journal will appreciate and welcome.

Boris S. Sokolov,
Editor-in-Chief

Stratigraphy and Geological Correlation, Vol. 1, No. 1, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 1, 1993, pp. 6 - 20.
Original Russian Edition Copyright © 1993 by Stratigrafiya. Geologicheskaya Korrelyatsiya, Semikhatov.
English Translation Copyright © 1993 by Interperiodica.

General Subdivision of the Precambrian: Comparison of the Latest Scales

M. A. Semikhatov

Geological Institute, Russian academy of Sciences, Pyzhevskii Per. 7, Moscow, 109017 Russia

Manuscript received January 20, 1992

Abstract – Conceptual principles, constructions, and substantiations of boundaries in three recently proposed Precambrian scales have been analyzed. These include the Scale of Precambrian time proposed by the International Subcommission on Precambrian Stratigraphy, the general scale of the Northern Eurasia Precambrian (within the former Soviet Union) adopted in 1990, and the Precambrian part of a geologic time scale proposed by W.B. Harland *et al.* in 1990. The first of the scales is based strictly on the chronometric concept. The second belongs to a chronostratigraphic category and meets the relevant requirements exactly, except in confining the boundaries to the bases of unconformity-bounded units in the type areas of corresponding boundaries and in determining their positions in the composite sequence of type areas. The third scale, which is a combination of chronostratigraphic and chronometric components, contains, along with several advantages, several inconsistencies in its construction. The prerogatives of national and international scales of the Precambrian are discussed, along with arguments favoring a general chronostratigraphic subdivision of the Precambrian, and suggestions for constructing a general chronostratigraphic scale.

Important events have recently occurred in the field of Precambrian stratigraphy: the International Stratigraphic Commission and International Union of Geological Sciences approved the Precambrian time scale proposed by the International Subcommission on Precambrian Stratigraphy in 1988 in Tianjin, the People's Republic of China (Plumb, 1991); the USSR Interdepartmental Stratigraphic Committee adopted a new stratigraphic scale of the Northern Eurasia Precambrian (within the former Soviet Union), that was accepted by the Second USSR Conference on the General Problems of Precambrian subvision in 1990 in Ufa (Semikhatov *et al.*, 1991); and a group of British scientists published a new version of a general Precambrian scale (Harland *et al.*, 1990). The first scale is of international status, the second reflects the experience and traditions of subdivision of the Precambrian in the vast expanses of Northern Eurasia, and the third attempts to unite the different approaches to constructing general Precambrian scales. There seems to be a need to compare the constructions and conceptual principles of these scales, and estimate their advantages and drawbacks. No such comparison exists, however, each of the scales was minutely discussed in literature (Plumb and James, 1986; Cowie *et al.*, 1989; Harland *et al.*, 1990; Plumb, 1991; Semikhatov *et al.*, 1991). Below are the main features of the three scales, and their comparative analysis.

I.

The scale of the International Subcommission on Precambrian Stratigraphy (SPS scale) and the scale accepted in Ufa are plotted in Table 1. In their construction, against the background of the coinciding time ranges and names of the largest Precambrian subdivisions, Archean and Proterozoic, there are at least five substantial differences. (1) In the SPS scale only the Proterozoic is subdivided into subordinate units, but in the USSR scale, both the Proterozoic and Archean are subdivided. (2) The Proterozoic in the SPS scale is subdivided into three units, but in the Northern Eurasia scale it is subdivided into two, though some Russian authors, such as V.E. Khain suggest dividing it into three units, in accordance with the subdivision adopted in the SPS scale, and such suggestions were submitted to the Ufa Conference (Akhmedov *et al.*, 1990; Rundkvist, 1990; Semikhatov *et al.*, 1990). (3) The Proterozoic in the compared scales is subdivided in varying detail (greater in the SPS scale), and the boundaries of subdivisions accepted in each scale usually do not coincide, although some, such as the Archean/Proterozoic boundary, do coincide. (4) Similar time range subdivisions in the two scales are considered to belong to different stratigraphic ranks. The three ranks recognized in the SPS scale are eon, era, and period. The Archean and Proterozoic are considered to be eons, and the units of 150 - 200 Ma time range are periods. The scale adopted in Ufa contains stratigraphic units of the following four subordinated ranks: Acrothem, Eonothem, Erathem, and system. In this scale, the Proterozoic and Archean are considered Acrothem, the units with the time ranges of hundreds of Ma are erathems, and the Vendian is recognized to be a Precambrian system. The Vendian time range is about 100 Ma (from 80 to 120 Ma depending on the admitted dating of the Cambrian lower boundary) and exceeds the time range of

the longest Phanerozoic period, the Ordovician, which is approximately 70 Ma. (5) The nomenclatures of the Proterozoic subdivisions in the two scales are different, and their differences, as it will be demonstrated below, are more than linguistic. The authors of the SPS scale consider it inadmissible to relate the names of the general Precambrian subdivisions to any geographical, ethnographical, or proper names of regional stratigraphic units of corresponding age. In the Northern Eurasia scale such a relation is considered necessary and is adopted for the entire system of the subordinate Proterozoic units (e.g., Karelian, Riphean, Karatavian, and Vendian). In contrast, the neutral names of eras, reflecting their position in the Proterozoic Eon, have been adopted in the SPS scale (Paleo-, Mesa-, and Neoproterozoic). The names of the periods were derived from Greek words meaning geological events which are considered typical of relevant periods, but do not determine them and are not strictly limited by their time boundaries.

The following events are regarded as suitable for naming the Proterozoic periods in the SPS scale: accumulation of iron formations for the Siderian (from sideros, iron), emplacement of Bushveld-type layered intrusions for the Rhyacian (from rhyax, lava flow), widely developed orogenic processes for the Orosirian (from orosira, mountain chain), stabilization of cratons for the Statherian (from statheros, stable, durable), formation of sedimentary covers of cratons for the Calymmian (from calymma, cover, coverlet), expansion of the cratonic sedimentary cover for the Ectasian (from ectasis, spreading, expansion), formation of the narrow zones of intense deformation and metamorphism for the Stenian (from stenos, narrow), development of the extension and rifting zones for the Tonian (from tonos, extension), and global glaciations for the Cryogenian (from cryos, ice). An exception has been made only for the youngest Proterozoic period, with the suggestion made to name it Neoproterozoic III until the International Working Group on the Terminal Precambrian System determines its name, stratotype, and lower boundary on the chronostratigraphic basis.

The above construction and terminology by no means represent all the differences between the scales under comparison. The scales are essentially different in that they represent the two contrasting concepts of the general Precambrian subdivision: the chronometric (SPS scale) and the chronostratigraphic (Ufa scale). These concepts which have been considered earlier (Semikhatov, 1991a), are summarized here.

The SPS scale, as with any chronometric scale, divides abstract geological time into intervals. The subdivisions and boundaries of these scales are defined by units of time, specifically years, on the basis of radiometric dating. The choice of suitable time units, which is purely theoretical, has been discussed by Harland *et al.* (1990), H.J. Hofmann (1991), and A.F. Trendall (1991). In its extreme expression, the chronometric concept requires construction of scales on the basis of arbitrarily chosen, equal time intervals (e.g., 100, 500 Ma, 10^{15} s, etc.), in no way related to the geological record (for bibliography and discussion see Semikhatov *et al.*, 1991a, and Hofmann, 1990, 1991). These scales are unpopular because of the geologists' need to search for geological information in natural rock successions and because of the very essence of general stratigraphic scales. Such scales should not only be the standards for the correlation of regional scales (which may be accomplished in principle by scales with arbitrary units of equal time range), but also the tools for reconstructing the succession of stratigraphic subdivisions and dividing the Earth's history into meaningful geological periods. Therefore, the authors of most chronometric scales, including the most popular (Harrison and Peterman, 1982; Stockwell, 1982; Plumb and James, 1986; Plumb, 1991) strive to relate the boundaries of chronometric units to some geological events. The SPS scale in this respect is no exception. Most of its subdivisions have a similar time range of 200 Ma (Table 1) because of a "fortunate" chronological arrangement of the events chosen for selecting the boundary positions in the scale, rather than the desire to use intervals of equal duration.

Before considering the specific events used in the SPS scale for selecting the positions of boundaries, it should be noted that these events were inferred and dated by analyzing the composite Precambrian geological record of the Earth, rather than being chosen, fixed, and dated in a single region or section (Plumb and James, 1986, p. 70). This approach, characteristic of many other Precambrian scales suggested during the last 20 to 25 years, differs from that of chronostratigraphic scales, where a boundary is determined by a certain event (or events) in a carefully selected section, and is fixed at a chosen and agreed point known as a "golden spike". The events used in the SPS scale for the selection of boundary positions are the major and most laterally persistent hiatuses in the deposition of supracrustal rocks and gaps in the plutonic and tectonic activity. The "boundaries will be selected so as to split as few known principle cycles of sedimentation, orogeny, and magmatism as possible" (Plumb and James, 1986) and to delimit or enclose considerable geological events.

A special reference concerning the actual gaps in plutonic activity is important for two reasons. First, modern, precise isotopic datings prove that in the early Precambrian, and especially in the Archean, stratified rocks represent only a very short interval of geological history. Its considerably lengthier stages are either represented by intrusive and deeply metamorphosed rocks, or not reflected in any rock complexes, and marked only by the disturbance of some isotopic geochronological system. Second, intrusive and deeply metamorphosed rocks relate to the main bulk of isotopic datings of today, so the statistic weight of such datings in determining chronometric boundaries becomes great. As a result, the crystalline complexes, which are difficult and often impossible to subdivide, become the elements determining the construction of the scale; whereas the genuine objects of stratigraphy, stratified

Table 1. General subdivision of the Precambrian: A – The SPS scale (Plumb, 1991); B – The Northern Eurasia scale (Semikhatov *et al.*, 1991). Figures indicate the isotopic age of boundaries, Ma

A			B			
Eon	Era	Period	Acrothem	Eonothem	Erathem	System
Cambrian base				Cambrian base		
Proterozoic	Neoproterozoic	Neoproterozoic III	Proterozoic	Upper		Vendian V_2 / V_1
		650		Riphean	650 ± 20	
		Cryogenian			Upper (Karatavian)	
		850				
		Tonian				
	1000				1000 ± 50	
	Mesoproterozoic	Stenian			Middle (Yurmatan)	
		1200				
		Ectasian			1350 ± 20	
		1400				
		Calymmian			Lower (Burzyanian)	
	1600			1650 ± 50		
	Paleoproterozoic	Statherian		Lower (Karelian)	Upper Karelian	
		1800				
		Orosirian			1900 ± 50	
		2050				
		Rhyacian			Lower Karelian	
		2300				
		Siderian				
2500			2500 ± 50			
Archean			Archean	Upper		
				3150 ± 50		
				Lower		

rocks, become secondary and sometimes are not considered at all in the construction of a scale. This tendency influences the opinion that crystalline complexes are either equally important as supracrustal rocks, or the focus of Precambrian stratigraphy (for bibliography see (Semikhatov, 1991a).

In considering the SPS scale, it should be emphasized that the analysis of the timing of main geological events is used only for the selection of the position of the subdivision boundaries. But the definitions of the boundaries are made only in units of isotopic age, in years, without relating the subdivisions and their boundaries to certain geological bodies and events.[1]

It is clear from the above that the units of the SPS scale do not and cannot have stratotypes. That is why any names of these units, implying a relation to certain geographic locality or a characteristic regional lithostratigraphic unit, do not correspond to the adopted concept that units of the scale should be completely independent of any natural geological bodies.

To illustrate the methods used in constructing the scale, consider the establishment of the time range and the inner subdivision of the unit that is now called Neoproterozoic, which began 1000 Ma ago, and was terminated by the beginning of the Cambrian. In the first version of the scale (Plumb and James, 1986), the beginning of this unit, then called Proterozoic III, was defined as 900 Ma, and its characteristic features were assumed to be as follows: (a) laterally widespread platform deposits (Upper Riphean of Northern Eurasia, Sinian of China, Adelaide of Australia, Infra-Cambrian of North Africa, etc.) closely structurally linked to Paleozoic strata and partly deformed together with them;

[1] The isotopic age of boundaries is given without an error factor, e.g., 1600 Ma rather than 1600 ± 50 Ma.

(b) traces of ancient glaciations; and (c) an appearance of the first Metazoa in its upper part. The manifestation and termination of the Grenville events, whose culmination was dated 1200 to 1000 Ma, and termination, 900 Ma, were considered typomorphic of the previous period concluding the Proterozoic II. Accordingly, the boundary between the Proterozoic II and Proterozoic III in the tentative version of the scale was defined as 900 Ma. It was proposed to divide the Proterozoic III into two periods with the boundary between them at 700 Ma. In substantiating such a subdivision and characterizing the first period of the Proterozoic III, K.A. Plumb and H.L. James (1986, p. 87) assumed that "the diagnostic feature of this period was the manifestation of several episodes of widespread continental glaciation in all tectonic regimes and on all continents," which were larger than any glaciation in the latest Proterozoic; the other feature was the beginning of the formation of an extensively distributed platform sedimentary cover. The culmination of the Pan-African-Brazilian orogeny and, more importantly, the appearance of multicellular Metazoa, the Ediacara fauna, were considered the characteristic features of the second period of the Proterozoic III (from 700 Ma ago to the beginning of the Cambrian).

However, at the Eighth Session of the International Subcommission on Precambrian Stratigraphy in 1988 at Tianjin, after analyzing the data in greater detail, it became clear that the chronology of the formation of widespread sedimentary covers on continental massifs in the end of the Proterozoic was more intricate, and that the three stages of this formation can be identified and clearly limited in time and space. The first stage, beginning about 1000 Ma ago, included the formation of sedimentary covers within the West African and East European cratons, and some paleogeographical rearrangements in the North China platform. Characteristic of the second stage, beginning about 850 Ma ago, was the formation of the Sinian in the South China platform, the Adelaide Supergroups in Australia, and their chronological analogues, which was preceded by significant geological events in many regions (Semikhatov, 1974; Khomentovskii, 1976, 1984; Borukaev, 1985; Khain and Bozhko, 1988; Research..., 1980). The third stage, beginning about 650 Ma ago, was characterized by the formation of the most widespread Precambrian sedimentary covers (Vendian and its analogues), and the appearance of Ediacara fauna (Sokolov, 1984, 1985). After analyzing the timing of these stages and revaluation of the age of the late Grenville events based on the new U-Pb zircon datings, the final version of subdivision of the Proterozoic terminal part (Table 1) was adopted by the Subcommission on Precambrian Stratigraphy. But the boundaries of adopted subdivisions, both in this and in the previous version of the SPS scale were "defined in years without specific references to any body of rock" (Plumb, 1991).

II.

The scale adopted in Ufa, unlike the SPS scale, belongs to the category of chronostratigraphic (chronostratic) scales (Semikhatov *et al.*, 1991) [2]. The scales of this category are based on the analysis of the natural succession, relationships, and characteristic features of material objects, rocks, and organic remains contained in them; establishing the stratotypes of subdivisions; and, importantly, on fixing the boundaries of stratigraphic units in specially selected type sections (boundary stratotypes), where the boundaries are determined by specific events and are formally fixed in a point marked by a golden spike (global stratotype section and point) (Cowie *et al.*, 1986; Cowie, 1989). Significance of the selection of a boundary stratotype depends on its role as a stratigraphic anchor, "to which all the subsequent correlations will be allocated, even if new paleobiologic or physical methods appear" (Holland, 1984, p. 70). The main criterion for fixing the boundary stratotype point is the widest and most reliable correlation potential of the relevant point by any or all methods available (Cowie *et al.*, 1986). An ideal general chronostratigraphic scale would have a succession of specially selected and internationally agreed upon points of boundary stratotypes, which would be determined by real events, having a good correlation potential, and separating named and ranked subdivisions.

Because of the characteristic features of the Precambrian rock record and the evolutionary level of its organic world, the boundaries in the Northern Eurasia Precambrian scale are determined by geological, rather than paleobiological events, as in a classical chronostratigraphic concept elaborated on Phanerozoic materials. And so the boundaries of general units in this scale, contrary to the Phanerozoic canons, have not been selected in continuous sequences, but were confined to the unconformities at the bases of the major stratified complexes, which marked the beginnings of important stages in the geological evolution of stratotype areas. This choice of events for determining boundaries, which became possible because of recent achievements in isotopic dating the stratified series, has been suggested before the conference in Ufa (Bibikova *et al.*, 1989; Semikhatov, 1991b) and was widely supported there (Semikhatov *et al.*, 1991). It was an important advance over the earlier version of the general Precambrian scale of the Soviet Union adopted in 1978 (Resolution ..., 1979; Semikhatov, 1979), in which the boundaries of major subdivisions were confined to the

[2] Chronostratigraphy, according to the International Stratigraphic Guide, (1976), is a part of stratigraphy dealing with the age of beds and their time relationships. This definition is sometimes currently used. However, the International Commission on Stratigraphy has specified the meaning of the term and regards chronostratigraphy as a part of stratigraphy "that defines a scale from continuous rock sequences with standardized reference points in carefully selected and agreed stratotype sections (Cowie *et al.*, 1984, p. 649). Harland *et al.*, (1982, 1990) define a chronostratigraphic scale simply as a scale of a rock succession with standardized boundaries selected in the boundary stratotypes.

terminations of important tectono-magmatic cycles, which inevitably increased the significance of plutonic rocks in determining the boundaries, with all the ensuing consequences (see above).

It is important to note one more feature of the scale adopted in Ufa: not only must the types (type lithostratigraphic complexes) of all adopted subdivisions be determined, as was done in the previous version of the scale (Resolution ..., 1979), but the next step must also be done. All its boundaries, except the boundary between the Lower and Upper Karelian, were confined to certain levels in the composite stratigraphic sections of stratotype areas (Semikhatov *et al.*, 1991). Such a confinement should be regarded as a first essential step in establishing the boundary stratotypes for these subdivisions, the series of especially selected points, each fixing a certain boundary in a single section.

It stands to reason that the type complexes of all subdivisions of the Northern Eurasia scale, embracing the entire Precambrian, and the candidates for its boundary stratotypes could not be selected in a single region. However, the natural properties of the Northern Eurasia Precambrian and the extent to which its regional complexes have been studied made it possible to select the type sections of the adopted units within the East European craton and its eastern surroundings, the South Urals, and to narrow the geographical localization of type sections for a number of successive subdivisions of the scale and for their boundaries even more. Such narrowing minimizes the danger of the type complexes of adjacent subdivisions being overlapped in time and favors the strict identification of their paleontological and other characteristics.

Thus, the geological events are used in the Northern Eurasia scale not only for the selection of boundary positions, as in the SPS scale, but also for defining these boundaries. As for the isotopic geochronological data, these are used in chronostratigraphic scales, including the scale under consideration, in two ways. First, they are used for dating (calibrating) the boundaries established by other methods (not for defining boundaries as in chronometric models). Such a calibration in the Northern Eurasia scale has been made either on the basis of the direct, mainly U-Pb zircon datings of the supracrustal rocks to whose base the relevant boundaries are confined, or by the interpolation between the U-Pb, Sm-Nd, and in the Upper Proterozoic, Rb-Sr datings of intrusive and supracrustal rocks bracketing a boundary in its type locality. Second, radiometric data are used for "projecting" the scale onto area, being one of the main means, and in the lower Precambrian the main tool of tracing subdivisions and their boundaries beyond their stratotype areas.

In this connection, it should be emphasized that each subdivision of, at least, the Proterozoic part of the scale adopted in Ufa reflects something much more important than the present concept of their time range. Their own, unique assemblages of organic remains, whose individuality and systematic diversity increase upsection, can be identified for each Upper Proterozoic unit and for the Karelian as a whole (Semikhatov *et al.*, 1990). The presence of these assemblages not only considerably enhances the correlation potential of the units, but also appreciably changes their determining essence. Moreover, the paleontological data in the scale adopted in Ufa are also involved in substantiating the subdivision. The twofold division of the youngest unit of the scale, the Vendian System, with the boundary along the base of the Redkino Horizon in the Moscow syneclise, has been based not only on a geological event, the beginning of the postglacial Valdai (Redkino) transgression, but also on a biotic event, the first wide lateral expansion and diversification of soft-bodied Metazoa. An alternative subdivision of the Vendian, with the boundary between the Red'kino and Kotlin horizons on the East European craton, is also based on ecostratigraphic data (Sokolov, 1984, 1985).

It seems likely that in the near future the biostratigraphic criteria will also play an appreciable role in subdividing (that is, in substantiating boundaries) deeper, pre-Vendian parts of the Proterozoic. The point is that the successive systematically different pre-Vendian assemblages of organic remains, though present discreetly in sections (including the type sections), exhibit the major renewals of their compositions in the lowermost parts of all the three Riphean erathems and the Vendian (stromatolites), the Lower, and especially the Upper Riphean (microfossils). This can be explained by the fact that the initial stages of the abovementioned units were marked by the highest appearance rates of new taxa among the relevant groups of Precambrian organisms.

It is essential that the specificity of the Precambrian chronostratigraphic units is determined or can be determined by more than just their paleontological contents. All the post-Early Riphean units, including the Vendian may already be characterized by a clearly specific carbon isotope signature, and some characteristic features of strontium isotopic composition can be outlined for post-Middle Riphean units. Current works are aimed at creating the paleomagnetic scale of the upper Precambrian, and elucidating the geochemical specialization of its particular parts. This may result in broadening the criteria for substantiating and correlating the units and, correspondingly, in enhancing the reliability and resolution of the scale. The possibility of such broadening is determined by the very essence of the chronostratigraphic concept, and is the main argument in favor of applying this concept to the subdivision of the Precambrian (for more detailed discussion and bibliography see Hedberg, 1974; Semikhatov, 1991a, b).

III.

The scale of Precambrian time proposed by Harland *et al.* (1990) is a development of a similar scale published earlier (Harland *et al.*, 1982). This scale has three distinctive features (Table 2).

First, its authors assume that for the Precambrian (as well as for the Phanerozoic) there is a need for two par-

Table 2. Precambrian linear time scale (Harland *et al.*, 1990)

Eons and subeons of the global chronometric scale		Chronostratigraphic scale			
		Eon	Era	Period	Epoch
		Phanerozoic	Paleozoic	Cambrian	
					570
Proterozoic	Pt_3	Precambrian	Sinian	Vendian	2 epochs
					610
				Sturtian	
					800
	900		Riphean	Karatau	
	Pt_2				1050
				Yurmatin	
					1350
				Burzyan	
	1600				1650
	Pt_1		Animikean	Gunflint	
					2200
			Huronian	I - IV	
	2500				2400 - 2500
Archean	Ar_3		Randian	V, VI VII	
					2800
	3000		Swazian	Pongola	
	Ar_2			Moodies Figtree Onverwacht	
	3500				3500
			Isuan		
					3800
	Ar_1		Hadean	Imbrian	
					3850
				Nectarian	
	4000				3950
				pre-Nectarian	
Priscoan					4150
				Moon and Earth's origin	
					4550 - 4570
			pre-Hadean		

Note. Epochs of the Hadean and most of the proposed symbols of subdivisions are omitted. Arabic numerals indicate the isotopic age of boundaries, Ma; Roman numerals, proper names of periods (from top downside): I – Cobalt, II – Quirke Lake, III – Hough Lake, IV – Elliot Lake, V – Ventersdorp, VI – Central Rand, VII – Dominion.

allel scales, a chronostratigraphic and a chronometric (Harland, 1975). The former scale should be based on rock successions, and should consist of intervals between certain events reflected in global stratotype sections and points. These intervals should also have proper names. The chronometric scale should be a linear time scale based on a standard time unit. The authors understand the chronometric scale in the widest sense, as any scale based on the units of duration, a standard second or a year. Therefore, it includes two categories of chronometric scales: (1) those which reflect the calibration of a chronostratigraphic scale in terms of isotopic age and are derived from chronostratigraphic scales and (2) those without any chronostratigraphic equivalents, whose boundaries are defined simply in years. The latter include completely artificial scales, which are independent of the geologic history, and consist of equal time intervals (e.g., Harland, 1975; Hofmann, 1990; Trendall, 1991), and scales, which reflect geologic events, and therefore, consist of variable time intervals (e.g., Harrison and Peterman, 1982; Plumb and James, 1986).

Second, the suggested scale includes not only the geological history of the Earth, but also the evolution of the Solar System and outer space, up to the Big Bang. Thus, the Precambrian part of the scale includes a large interval not represented in the rock record of the Earth.

Third, the geological time scale is regarded by its authors as consisting of chronostratigraphic and chronometric components, and both these components have been considered in detail for the Precambrian. Each will be discussed here.

The chronometric component of the scale is simple: it identifies the three eons, Priscoan, Archean, and Proterozoic, and subdivides the Archean and Proterozoic into subeons. The first eon of the scale embraces the entire pre-Archean history of the Solar System; a round number of 4000 Ma was arbitrarily chosen as the Priscoan/Archean boundary, and the Archean was arbitrarily divided into three intervals of 500 Ma each. In defining the beginning of the Proterozoic and its subdivisions, Harland *et al.* followed an early version of the SPS scale (Plumb and James, 1986), reflecting real events imprinted in the composite geological record of the Earth. Thus, for constructing different parts of the Precambrian chronometric scale, Harland *et al.* used different approaches. They identified arbitrary, equal intervals unrelated to the real Earth's history, but also identified variable intervals reflecting significant events in geological history. They unambiguously interpreted the Priscoan and Archean eons as purely chronometric subdivisions, and regarded the Proterozoic either as a chronometric unit (Harland *et al.*, 1990, pp. 13-18) or as a unit of dual, chronometric and chronostratigraphic belonging (Harland *et al.*, 1990, p. 2).

The chronostratigraphic component of the scale under consideration does not formally pretend to be of international significance, and is regarded by the authors as an example of elaborating the Precambrian chronostratigraphic scale in the presence of good sections. It consists of a system of subordinated units of several ranks: eon, era, period, epoch, and, for the uppermost part of the Precambrian, a stage. The entire Precambrian, contrasted with the Phanerozoic, is considered to be an eon; it is divided into nine eras with a duration of 800 - 200 Ma, which fills the entire scale. Periods with duration of 400 Ma or less have been identified, but mostly only outlined within the eras, and more minute subdivisions have been defined for some intervals of the scale.

Thus, a chronostratigraphic part of the geological time scale suggested by Harland *et al.* (1990) has, as a whole, a greater resolution than its chronometric component, but is also inconsistent in the construction of its different parts.

The first subdivision of the chronostratigraphic scale, the pre-Hadean, is considered to be an era in rank, but should not, strictly speaking, appear in a scale of this category. It has no material expression whatever, and it precedes the Moon and Earth's origin (4.55 - 4.57 Ga ago), so, at best, it could pretend to be a chronometric unit. The second subdivision of the same rank, the Hadean, is considered to correspond to the pre-Archean Earth's history, but its material expression is assumed to be the oldest lunar rocks, and the superposition of lunar craters and seas is used for subdividing the Hadean into subordinate units, whose boundaries have been dated on the basis of the radiometric datings of extraterrestrial objects. At the same time, Harland *et al.* (1990, p. 22) have noted that the Hadean "initial boundary coincides with some event taken as signal to the first step in the formation of the Earth." It is clear that this boundary cannot meet the requirements for the boundaries of chronostratigraphic units, on whose compulsory execution insist Harland *et al.* Therefore, this boundary and the entire Hadean in its suggested time range should not be regarded as chronostratigraphic objects.

The younger, post-Hadean eras have been more strictly defined, however, some contradictions and inconsistencies also exist in their definitions. A succession of the best known and well studied regional lithostratigraphic complexes was selected for determining the types of these eras. These unconformity-bounded complexes represent considerable intervals of the geological history, and most of them have already been adopted as the types of general subdivisions in other scales. Thus, the Witwatersrand and Ventersdorp supergroups of South Africa have been accepted as the type of the Randian Era, and the succession of the Burzyan, Yurmata, and Karatau groups of the Urals, except the uppermost part of the latter, as the type of the Riphean Era, and so on. In these cases, the age relationships of the complexes situated far from one another have been defined mainly by radiometric methods, and in the upper part of the scale, partly by paleontological data.

In considering the post-Hadean Precambrian eras, the authors of the scale do not define and, with a single exception, do not suggest defining the exact position of their boundaries in particular or even in composite sec-

tions, although defining the chronostratigraphic units by means of selecting and fixing stratotype sections and points of their lower boundaries was emphasized in their work. Thus, only the possible geographical position of the Sinian and Riphean boundary stratotypes has been discussed, and it has been mentioned only briefly that the lower boundaries of other eras in the geological stage of the Earth's history (more exactly, their initial boundaries, as these are geochronological units) may be selected near the bases, but not at the bases of the type complexes of relevant subdivisions. Such a disengagement of the boundaries from the bases of unconformity-bounded complexes seems to be a logical step in a classical chronostratigraphic model. However, this step requires indicating the events that should, or at least could define the position of the Precambrian scale boundaries in their stratotypes. Unfortunately, such an indication has been made (as one of the possible alternatives) only with respect to the initial boundary of the Randian (the beginning of the tillite accumulation prior to the Still Water Complex, Montana, USA), and it is possible to conclude from the context of the work that the beginning of the Sinian may be defined by the beginning of the Sturtian Glaciation in southern Australia.

With respect to the subdivisions of lower ranks in the scale under consideration, only the part of the scale which is related to the geological stage of the Earth's history will be discussed. In the first eras of this stage, preceding the Riphean, the subordinate subdivisions (periods) are only outlined, their boundaries are not discussed and, with a single exception, are not drawn. Only from Figs. 7, 8, in the treatise by Harland *et al.*, can it be inferred that the prototypes of the periods outlined are conceived to be the successive groups of the supergroups forming the type subdivisions of relevant eras (Table 2).

As for the Riphean and Sinian eras, their subdivision into periods and, for the late Sinian, into epochs and stages (Harland *et al.*, 1990, Fig. 3.1) was discussed and adopted. The authors consider it necessary to adopt for the Riphean part of the chronostratigraphic scale the Ural model of its subdivision into the Early, Middle, and Late Riphean (Burzyan, Yurmatin, and Karatau periods), somewhat decreasing the Karatau time range at the expense of assigning its terminal part to the Sinian. The two periods, Sturtian and Vendian, have been identified within the Sinian, and the Sturtian and Varangerian (Laplandian) glaciations respectively, have been discussed as indicators of their initial boundaries. Once again, it was stressed that the initial boundary of the Vendian has to be fixed within the Varanger Glacial Horizon, rather than at its base. This departure of the Vendian time range (Harland *et al.*, 1990) from its original magnitude (Sokolov, 1984, 1985), being small, is significant from a conceptual standpoint, because it reflects two approaches to defining the Vendian lower boundary: "somewhere near the lowest glacial facies" in the conformable strata of the Varanger Horizon (Harland *et al.*, 1990, p. 30) or along the inconformity at the base of the lowermost (Blon' or Norway) tillite level of this horizon (Sokolov, 1984, 1985; Bibikova *et al.*, 1989; Semikhatov *et al.*, 1991).

The Vendian Period adopted in such a time range is divided into the Varanger and Ediacara epochs, and the boundary between them is selected somewhere in the lower part of the Wonoka Formation of Australia (Harland *et al.*, 1990). Therefore, the time range of the Ediacara Epoch differs somewhat from that of the Ediacaran Period identified earlier by R.J.F. Jenkins, (1981) with the lower boundary in the stratotype confined to the unconformity at the base of the Wonoka Formation.

The authors of the scale suggest subdividing the Vendian into even greater detail, and propose to delimit two stages in each of the identified epochs. In this case, the criteria of substantiation of such minute for the Precambrian subdivisions, and the levels of drawing their boundaries are not considered. However, the proposed names of these subdivisions, derived from two successive stages of the Varangerian Glaciation in Northern Norway and two upper formations of the Wilpena Group in southern Australia, suggest that it is the sections of these regions that are considered most important for substantiating the assumed Vendian subdivision.

The aforementioned scale of the upper horizons of the Precambrian (of the Sinian) is the most detailed scale ever suggested for this part of the geological record. However, the possibility of wide correlations of its minute subdivisions will be rather unclear, even if their boundaries are strictly defined and contradictions in estimating the radiometric age of the Cambrian lower boundary are successfully solved, while the adopted procedure of "assembling" the scale by selecting stratotypes of successive periods and epochs and their boundaries in distant sections creates a number of difficulties, which cannot always be solved with the current methodological basis of Precambrian stratigraphy. This is illustrated by a dual interpretation of the stratigraphic position of the Nantuo Tillite by the authors. This tillite is the key object of the scale under consideration, and in the type section of the Sinian in China, in the Yangtze Gorges, it constitutes the second from the base formation of the Sinian. First, the authors (Harland *et al.*, 1990, p. 28) consider it to be a probable member of the Sturtian (i.e., pre-Vendian) Glaciation and then (p. 30), a member of the Varangerian (i.e., early Vendian) Glaciation. It is clear that, depending on the correlation of the Nantuo Tillite, the position of the type of the Sinian (the real subdivision naming the terminal Precambrian era in the scale compiled by Harland *et al.*) changes, and the question arises concerning the correspondence between the Sinian time ranges in the proposed interpretation and the one adopted by Chinese geologists (in the Precambrian scale of China, the lower boundary of the Sinian is drawn along the inconformity at the base of the Liantuo Formation, which is disconformably overlain by the Nantuo Tillite in the

Yangtze Gorges (Research..., 1980; Keller *et al.*, 1984; Sun Dazhong and Lu Songnian, 1985)).

Harland *et al.* (1990) do not specifically dwell on the substantiation of isotopic ages accepted by them for the initial boundaries of the Sinian and pre-Riphean subdivisions (Table 2), and in the Riphean part of their scale follow the dates adopted in the USSR Precambrian scale of 1978 (Resolution..., 1979); though they depart from the estimated isotopic age of the Vendian lower boundary (610 Ma instead of 650 ± 20 or 650 ± 10 Ma) adopted in this scale and other publications (e.g., Sokolov, 1984, 1985). The departure was caused by the appearance of new U-Pb datings of the late Precambrian rocks in Newfoundland, but Harland *et al.* (1990) did not consider their correlation with Vendian-type sections and with the probable stratotype of the Vendian lower boundary.

Concluding the analysis of the geological time scale proposed by Harland *et al.*, the following remarks should be done.

First, the time resolution of the chronostratigraphic part of the scale is considerably greater than that of the chronometric part and does not depend on age. This independence is illustrated by the absence of directional upsection changes in the time ranges of the units of the same rank and by identifying the subdivisions of the fourth rank, epochs, only in the lowermost (Hadean) and uppermost (Vendian) intervals of the chronostratigraphic scale.

Second, the isotopic ages of the boundaries in the chronostratigraphic and chronometric components of the scale coincide only in a single case (the Isuan/Swazian boundary and the Early/Middle Archean) and approximately coincide in two other cases, whereas in most cases, the boundaries in these components are drawn at different time levels (Table 2). This fact emphasizes the independence of the adopted chronostratigraphic and chronometric subdivisions.

Third, there is no correlation between the time ranges of subdivisions and their ranks in the chronostratigraphic scale. For instance, the time range of eras varies from 230 to 850 Ma, and that of periods from 40 to 300 Ma. As a result, a question naturally arises concerning the criteria of subdivision ranks. For example, why the Huronian and Sinian with time ranges of 200 - 300 and 230 Ma respectively, are assigned to eras, and the Burzyan and Vendian with time ranges of 300 and 40 Ma respectively, both are considered to be periods?

Fourth, the type sections of the successive Precambrian subdivisions and their tentative boundary stratotypes were often selected on different continents. Such a selection, to a considerable extent dictated by the desire to introduce the most well-known subdivisions of various national scales into the scale under consideration, is commonplace for Phanerozoic stratigraphy, but its application to the Precambrian creates considerable correlation difficulties, which especially affect the correct identification of the subdivisions of the lowest ranks. One example was considered above.

Fifth, a number of traditional stratigraphic subdivisions (Riphean, Karatavian, Sinian, Vendian, etc.) in the scale are treated, in terms of time ranges, somewhat differently from usually adopted ones. This change of time ranges, partly connected with the displacement of initial boundaries of the subdivisions from the inconformities at the bases of the type groups within these groups, only in one case was terminologically reflected (Ediacara Epoch and Ediacaran Period). This results in dual meanings for time ranges of units of the same name in different scales. In most cases, this duality is significant only from a conceptual, rather than from a pragmatic standpoint, but with respect to the Riphean and Karatavian (Upper Riphean), it affects the very essence of these stratigraphic units: traditionally, their upper boundaries have been determined by the Vendian, rather than the Sinian, lower boundary.

IV.

So, of the three latest scales of general subdivisions of the Precambrian considered above, the geological time scale proposed by the International Subcommission on Precambrian Stratigraphy, which is, by definition, an international (global) scale, follows the adopted concept of construction most strictly. A single deviation from the concept assumed in the scale is the recognition of the possibility of defining the youngest period of the Neoproterozoic chronostratigraphically, which was reflected in its name.

The Precambrian scale, suggested by Harland *et al.* (1990), was supposedly globally significant, yet it proved to be more vulnerable from the standpoint of inner consistency and constructional strictness. Its chronometric part is fairly strict in its approach to boundary definition, but positions of the boundaries were selected in different ways for its different intervals, and two contradictions exist in its chronostratigraphic part. First, a considerable interval concerns the time which is not represented by the Earth's rock record and therefore, may not be considered in a chronostratigraphic context. Second, in the interval represented by this record, the scale, based on the succession of natural supracrustal rock complexes used for strictly defining stratigraphic units, is devoid of another important component of chronostratigraphic concept, the fixation of subdivision boundaries in the specially selected stratotype sections. The authors only note that a boundary may be selected near the base of a formation or group, but formally fix neither the exact position of the boundaries nor the events determining them.

A requirement to select the boundary stratotypes within conformable sequences, perceived from Phanerozoic stratigraphy, was advanced as applied to the Precambrian previously (e.g., Hedberg, 1974) and was put to use in the South African Precambrian scale suggested in the late 1970s (Kent, Hugo, 1978; South African..., 1990). However, an uncertainty concerning the

events that could define selected boundaries has led South African geologists to other opinions. It was suggested to allocate the boundaries of general subdivisions to the most important gaps in the sedimentation, magmatism, orogenesis, and cratonization processes (Johnson *et al.*, 1989), or to the inconformities at the base of the supracrustal rock type sequences (Winter, 1990). This last method of defining (fixing) the boundaries was adopted in other chronostratigraphic scales of the Precambrian or upper Precambrian (e.g., Research..., 1980; Sun Dazhong and Lu Songnian, 1985). In this case, the boundaries are assumed to be isochronous planes, whose place is determined by the position of inconformity in a selected type section, rather than the inconformities proper, which usually are diachronous.

The scale, referring to the Precambrian of Northern Eurasia (within the limits of the former Soviet Union) and meant for its subdivision and correlation, does not formally claim to be of a wider significance. However, the subdivisions of its Upper Proterozoic part, the Riphean with its subordinate units and the Vendian, have, in fact, become internationally current, which is specifically reflected in the introduction of these subdivisions into the scale suggested by Harland *et al.*, (1990); and the Lower Proterozoic (Karelian) of the Northern Eurasia scale corresponds in its time range to the Paleoproterozoic of the SPS scale based on the analysis of world-wide data. This global aspect and fairly strict adherence to a chronostratigraphical principle of construction put the USSR scale of 1990 into a special place among the other scales officially adopted in countries with the representative and widely distributed Precambrian (e.g., Kent and Hugo, 1978; Research..., 1980; Harrison and Peterman, 1982; Stockwell, 1982; Sun Dazhong and Lu Songnian, 1985; Johnson *et al.*, 1989; South African..., 1990).

The use of a chronostratigraphic principle in the Precambrian scales, as was said above and noted in the literature, has its own characteristic features related to the fact that, with the extent to which Precambrian organic remains are now studied, they cannot be used as a tool for typifying and tracing the stratigraphic boundaries interregionally. And so, the boundaries are determined in their stratotype localities by abiotic (geological) events, and maintain some correlation between the significance of an event and the rank of a corresponding boundary.

The scale adopted at Ufa in 1990 was compiled that way. All its boundaries, except the Lower/Upper Karelian boundary, were determined in their type areas by the beginning of the deposition of volcano-sedimentary or sedimentary complexes, marking the commencement of the important evolutionary stages of these areas and separated from underlying formations by the inconformities of some particular significance. The confinement of the boundaries to real events recorded at the base of inconformity-bounded type sequences, rather than within conformably superimposed sequences, and the formal definition of the boundary positions in the composite sections of type areas rather than in specific stratotype sections and points, cover all the significant deviations of this scale from the requirements of chronostratigraphy. It should be noted that currently, the general Phanerozoic stratigraphic scale does not meet all these requirements either. Therefore, it is only the degree of deviation from the requirements that matters in the problem.

The problem of the stratigraphic ranks of identified units is solved in the Northern Eurasia scale rather strictly, and this is its advantage over a number of other Precambrian scales. In this case, the time range of subdivisions defined by radiometric dating served as a criterion (Bekker, 1990, 1992; Semikhatov *et al.*, 1991). Substantial differences in the time ranges of the units of different ranks has long been a topic of discussion in literature and was especially stressed by V.V. Menner (1977, 1979). However, applying this criterion to the definition of the rank of Precambrian stratigraphic units was hampered in Russian literature by the influential opinion that the particular units, phytems or protosystems, differing from those applied for the Phanerozoic, should be identified in the Precambrian (Keller, 1966, 1980). This point of view, which I once shared, assumes that the methods of general subdivision of the Precambrian are specific, that its general units are substantiated by organic-sedimentary structures rather than by organic remains proper and do not form a multistage hierarchy of subordinated units typical of Phanerozoic systems. The latest understanding of the principles of identifying Precambrian chronostratigraphic units, the substantiation of the Vendian as the Terminal Precambrian System, the suggested identification of subordinate general subdivisions in the Upper Riphean, and the more precise definition of the time ranges of various Archean and Proterozoic subdivisions in Northern Eurasia – all this made it possible to apply the same criteria to define stratigraphic rank of both Phanerozoic and Precambrian subdivisions, and to form a hierarchy for the system of units adopted in the Northern Eurasia scale (Table 1).

Referring to the comparison of this and the SPS scales, note that if the time positions of the entire boundary successions of these scales are different, the ages of four pairs of the boundaries are actually the same. These are the boundaries with the ages of 2500 ± 50 and 2500 Ma, 1650 ± 50 and 1600 Ma, 1000 ± 50 and 1000 Ma, and 650 ± 20 and 650 Ma respectively. The assumed interpretation of the significance (place in a hierarchical order) of the first and second of the pairs also coincide. One of them is considered to be the Archean/Proterozoic boundary in both scales, and the other, the boundary between the first largest subdivision of the Proterozoic (named the Karelian in the Northern Eurasia scale, and Paleoproterozoic in the SPS scale) and the following subdivision of the same rank. There is no such coincidence in the interpretation of the significance of the third and fourth pairs of the above boundaries.

The boundary at 1000 Ma in the Northern Eurasia scale is a boundary between third-rank units subordinated to the Upper Proterozoic (Middle and Upper Riphean), whereas in the SPS scale it is the boundary between the second-rank subdivisions (Mesa- and Neoproterozoic), appearing in the same "weight class" as the 1600 Ma boundary and determining, together with the latter, threefold subdivision of the Proterozoic. As for the boundary at 650 Ma, in the scale adopted in Ufa, it is the lower boundary of a single fourth-rank subdivision, the Vendian, and in the SPS scale, it ranks with the boundaries subordinated to the Paleo-, Mesa-, and Neoproterozoic, i.e., it appears as a third-rank boundary.

The threefold subdivision of the Proterozoic adopted in the SPS scale, to some extent coinciding with those suggested in some other scales (e.g., Harrison and Peterman, 1982; Stockwell, 1982) and supported by a number of participants at the Ufa Conference, is rapidly becoming adopted in current literature. It contrasts with the twofold subdivision of the Proterozoic adopted at Ufa in 1990 and in the USSR Precambrian scale of 1978 (Resolution..., 1979; Semikhatov, 1979). Analyzing the ca. 1000 Ma-old events in this context, one easily finds that they were most strikingly expressed in a dynamothermal reworking of older crystalline complexes and in accompanying intrusions; while their manifestations in stratified rocks, sometimes attracting attention by their scale (e.g., the Late Riphean transgression in North Africa), do not virtually stand out against the background of other Late Proterozoic events (Semikhatov, 1974; Borukaev, 1985). Moreover, the events under consideration tended to be restricted to the present-day peri-Atlantic regions, whereas in the peri-Pacific and Central and Northern Asia they were, if manifested at all, very slack and clearly less significant than the 800 - 850 Ma-old events (Semikhatov, 1974; Khomentovskii, 1976; Khain, Bozhko, 1988). Therefore, Chinese and Siberian events were used for substantiating on the basis of the geological history stratigraphic boundaries at ca. 850 Ma ago, the lower boundaries of the Sinian and Baikalian (Khomentovskii, 1976, 1984; Research..., 1980, Sun Dazhong, Lu Songnian, 1985), because tracing the 1000 Ma-old events in the Pacific margins of North America and in Siberia is a rather difficult task, and not always unambiquously solvable.

In summary, the twofold subdivision of the Proterozoic adopted in the Northern Eurasia scale reflects the significance of the pre-Riphean events and the geological unity of the Riphean in Northern Eurasia, with its three subdivisions corresponding to the three successive stages in forming a platform cover of epi-Svecofennian cratons and of the destruction of the cratons which was confined mainly to the future Phanerozoic fold belts (Borukaev, 1985; Khain, Bozhko, 1988; Kiselev, 1991).

Any chronostratigraphic scale reflects in its subdivisions events of various scope of manifestation, and this explains the presence of both coinciding and noncoinciding boundaries in various scales, as in the scale adopted in Ufa, the SPS scale, and the chronostratigraphic Precambrian scales of China and South Africa (Kent, Hugo, 1978; Research..., 1980; Stockwell, 1982; Johnson *et al.*, 1989; South African..., 1990). For example, the Riphean lower boundary, determining the main Proterozoic subdivisions in the Northern Eurasia scale and coinciding with the Paleoproterozoic/Mezoproterozoic boundary in the SPS scale, in the Chinese and South African scales falls within such units as the Changcheng System and Mocolian Erathem respectively, and the lower boundary of the Vendian cuts the most prestigious Precambrian unit of China, the Sinian System.

Such examples are numerous, but the aforementioned sufficiently illustrates the specific character of the constructions of scales based on the analysis of geological history of distant regions with the representative Precambrian sequences. The national Precambrian scales, reflecting specific features of geological history of large segments of the Earth, will probably be retained for a long time. Each of them will ensure correlation of regional schematics and reconstruction of continuous successions of stratigraphic units as well as reflect the division of the Precambrian history into geologically meaningful periods within its segment. Attempts to combine some relatively small intervals of several such scales and to assemble them into a united chronostratigraphic Precambrian scale (Harland *et al.*, 1982, 1990) require, as we have seen, a considerable improvement and do not seem to be the most effective in their very essence.

The choice of the longest ranged natural stratigraphic successions, that because of their natural features and degree of understanding could become the types of two or three largest intervals of the scale, seem to be a methodologically more justified and more effective way of creating an international chronostratigraphic Precambrian scale. The list of claimers to such a choice is very limited. For the lower Precambrian, these are the successions from the southern Canadian shield and from South Africa, and for the upper Precambrian, those from China and particularly from Russia. It is not by chance that just these successions have served as models for the elaboration of well-known Precambrian scales.

With the presence of several national Precambrian scales officially approved in corresponding countries and differing in time ranges, names, ranking of adopted subdivisions, as well as in construction principles, an instrument for their correlation is required, which would be a common denominator for coexisting scales. The SPS scale (particularly, its largest subdivisions) will probably accomplish these functions, ensuring the mutual understanding of geologists adhering to different schools and scales. It is clear that correlation with the SPS scale may be executed only on the basis of isotopic geochronological data.

However, another approach to correlation and possible integration of various chronostratigraphic Pre-

cambrian scales seems to be more promising. It is based on extending the comprehensive characteristics of stratigraphic units in each scale and using the most evidence for their substantiation, specifying the time ranges, and, mainly, for correlation. Only a composite of features characterizing a unit makes it possible to use one of the main principles of stratigraphy, a principle of chronological interchangeability of features, which "makes it possible to combine features and to establish the chronological relations of spatially dissociated bodies with different stratigraphic features" (Meyen, 1989). This is why the use of multidisciplinary (biological, geological, climatological, physical, geochemical, etc.) criteria for substantiating the high-resolution stratigraphic scales recently became the main line in developing Phanerozoic stratigraphy. This line, which has been named holostratigraphy (Walliser, 1986; Cowie, 1989), is most successfully developed on Phanerozoic material. However, there are even more reasons to fully apply it to the Precambrian, too. Of course, this can be done only in the chronostratigraphic scope.

REFERENCES

Akhmedov, A.M., Bogdanov, Yu.B., Konikov, A.Z., *et al.*, Principles and Practice of Subdivision and Correlation of the Lower Precambrian of the Soviet Union, *Obshchie Voprosy Raschleneniya Dokembriya SSSR. Tezisy Dokladov 2-go Vsesoyuznogo Soveshchaniya* (General Problems of Subdivision of the Precambrian in the Soviet Union, 2nd National Conference), Abstracts of Papers, Ufa, 1990, pp. 3 - 5.

Bekker, Yu.R., A Rank of General Precambrian Stratigraphic Units, Obshchie Voprosy Raschleneniya Dokembriya SSSR. *Tezisy Dokladov 2-go Vsesoyuznogo Soveshchaniya* (General Problems of Subdivision of the Precambrian in the Soviet Union, 2nd National Conference), Abstracts of Papers, Ufa, 1990, pp. 5 - 7.

Bekker, Yu.R., A Rank of General Stratigraphic Units of the Precambrian Scale, *Izvestiya RAN. Ser. Geol.*, 1992, no. 9, pp. 5 - 10.

Bibikova, E.V., Lobach-Zhuchenko, S.B., Semikhatov, M.A., *et al.*, Precambrian Geochronological Scale of the East European Craton and Its Surroundings, *Izvestiya AN SSSR. Ser. Geol.*, 1989, no. 4, pp. 8 - 22.

Borukaev, Ch.B., *Struktura Dokembriya i Tektonika Plit* (Structure of the Precambrian and Plate Tectonics), Nauka, Novosibirsk, 1985.

Cowie, J.W., Natural Boundaries in the Global Stratigraphic Time Scale of IUGS, *28th Intern. Geol. Congress,* Abstracts of Papers, Washington, 1989, vol. 1, p. 336.

Cowie, J.W., Ziegler, W., and Remane, J., Stratigraphic Commission Accelerates Progress, 1984 to 1989, *Episodes*, 1989, vol. 12, no. 2, pp, 79 - 81.

Cowie, J.W., Ziegler, W., Boucout, A.J., *et al.*, *Guidelines and Status of the International Commission on Stratigraphy,* Cour. Forsch. Inst., Senckenberg, 1986.

Cowie, J,W., Basset, M.G., Fritz, W.H., *et al.*, Chronostratigraphic Definition of the Precambrian-Cambrian Boundary, *Geol. Mag.*, 1984, vol. 121, no. 6, p. 649.

Harland, W.B., The Two Geological Time Scales, *Nature*, 1975, vol. 253, no. 5492, pp. 505 - 507.

Harland, W.B., Armstrong, R.,L., Cox, A.V., *et al.*, *A Geologic Time Scale 1989*, Cambridge University, Cambridge, 1990.

Harland, W.B., Cox, A.V., Llewellyn, P.G., *et al.*, *A Geologic Time Scale,* Cambridge University, Cambridge, 1982.

Harrison, J.E. and Peterman, Z.E., North American Commission on Stratigraphic Nomenclature. Report 9. Adoption of Geochronometric Units for Divisions of Precambrian Time, *Bull. Amer. Assoc. Petrol. Geol.*, 1982, vol. 6, no. 6, pp. 801 - 802.

Hedberg, H.D., Basis for Chronostratigraphic Classification of the Precambrian, *Precambr. Res.*, 1974, vol. 1, no. 1, pp. 165 - 177.

Hofmann, H.J., Precambrian Time Units and Nomenclature – the Geon Concept, *Geology*, 1990, vol. 18, no. 2, pp. 304 - 341.

Hofmann, H.J., Precambrian Time Units – Geon or Geologic Unit?, *Geology*, 1991, vol. 19, no. 9, pp. 958 - 959.

Holland, C.H., Steps on the Way to the Standard Silurian, *27 MGK. Stratigrafiya. Sektsiya C.01. Doklady* (27th Intl. Geol. Congress. Stratigraphy. Section C.01. Reports), Moscow, 1984, vol. 1, pp. 60 - 72.

International Stratigraphic Guide, Hedberg, H.D., Ed., John Wiley and Son, New York-London-Sydney-Toronto, 1976.

Jenkins, J.F., The Concept of an "Ediacaran Period" and Its Stratigraphic Significance in Australia, *Trans. R. Soc. S. Austral.*, 1981, vol. 105, no. 4, pp. 179 - 194.

Johnson, M.R., Cornell, D.H., and Walraven, F., A Revised Precambrian Time Scale for South Africa, S. Afric. Commission on Stratigraphy. *Chronostrat. Ser. 1. Geol. Surv. S. Africa*, 1989.

Keller, B.M., Subdivisions of a General Stratigraphic Precambrian Scale, *Doklady AN SSSR*, 1966, vol. 171, no. 6, pp. 1405 - 1408.

Keller, B.M., *Stratigraficheskaya Klassifikatsiya. Materialy k Probleme*, (Stratigraphic Subdivisions of the Precambrian), Nauka, Leningrad, 1980, pp. 116 - 124.

Keller, B.M., Semikhatov, M.A., Chumakov, N.M., Type Sections of the Proterozoic Upper Erathem, *27 MGK. Geologiya Dokembriya. Sektsiya C.05. Doklady* (27th Intern. Geol. Congress. Geology of the Precambrian. Section C.05. Reports), Moscow, 1984, vol. 5, pp. 56 -76.

Kent, L.E., and Hugo, P.J., Aspects of the Revised South African Stratigraphic Classification and a Proposal for Chronostratigraphic Subdivision of the Precambrian. Contributions to the Geologic Time Scale, *Amer. Assoc. Petrol. Geol.*, Tulsa, 1978.

Khain, V.E., and Bozhko, N.A., *Istoricheskaya Geotectonika. Dokembrii* (Historical Geotectonics. Precambrian), Nedra, Moscow, 1988.

Khomentovskii, V.V., *Vend* (The Vendian), Nauka, Novosibirsk, 1976.

Khomentovskii, V.V., Baikalian Complex and Baikalides, *Geol. i Geofiz.*, 1984, no. 8, pp. 33 - 40.

Kiselev, V.V., *Dokembrii Paleozoid Tyan' Shan'* (Precambrian of the Tien Shan Paleozoides), Doctorate (Geol.-Min.) Dissertation, Bishkek, 1991.

Meyen, S.V., *Vvedenie v Teoriyu Stratigrafii* (Introduction to the Stratigraphic Theory), Nauka, Moscow, 1989.

Menner, V.V., A General Scale of Stratigraphic Subdivisions, *Izvestiya AN SSSR, Ser. Geol.*, 1977, no. 11, pp. 8 - 15.

Menner, V.V., A General Scale of Stratigraphic Categories in the Light of Last Results of Geological Works, *Byul. MOIP. Otd. Geol.*, 1979, vol. 54, no. 2, pp. 31 - 48.

Plumb, K.A., New Precambrian Time Scale, *Episodes*, 1991, vol. 14, no. 2, pp. 139 - 140.

Plumb, K.A. and James, H.L., Subdivision of Precambrian Time: Recommendations and Suggestions by the Subcommission on Precambrian Stratigraphy, *Precambr. Res.*, 1986, vol. 32, no. 1, pp. 65 - 92.

Research on Precambrian Geology: Sinian Suberathem in China, Sci. Technol., Tianjin, 1980.

Resolution of the USSR Conference on General Problems of Subdivision of the Precambrian in the USSR, *Obshchie Voprosy Raschleneniya Dokembriya SSSR*, Nauka, Leningrad, 1979, pp. 147 - 153.

Rundkvist, D.V., On Principles of Elaboration of a Lithostructural Scale of Geological Time, *Obshchie Voprosy Raschleneniya Dokembriya SSSR. Tezisy Dokladov 2-go Vsesoyuznogo Soveshchaniya* (General Problems of Subdivision of the Precambrian in the Soviet Union, 2nd National Conference), Abstracts of Papers, Ufa, 1990, pp. 24 - 29.

Semikhatov, M.A., *Stratigrafiya i Geokhronologiya Proterozoya* (Stratigraphy and Geochronology of the Proterozoic), Nauka, Moscow, 1974, p. 302.

Semikhatov, M.A., New Stratigraphic Scale of the USSR Precambrian: Analysis and Lessons, *Izvestiya AN SSSR, Ser. Geol.*, 1979, no. 11, pp. 5 - 22.

Semikhatov, M.A., Modern Concepts of General Subdivision of the Precambrian: an Analysis, *Izvestiya AN SSSR, Ser. Geol.*, 1991a, no. 8, pp. 3 - 13.

Semikhatov, M.A., General Problems of Proterozoic Stratigraphy in the USSR, *Soviet Scientific Rev., Section B. Geology Reviews*, Harwood Academy, 1991b.

Semikhatov, M.A., Fedonkin, M.A., Veis, A.F., *et al.*, Paleontological Method in Precambrian Stratigraphy, Obshchie Voprosy Rascleneniya Dokembriya SSSR. *Tezisy Dokladov 2-go Vsesoyuznogo Soveshchaniya* (General Problems of Subdivision of the Precambrian in the Soviet Union, 2nd National Conference), Abstracts of Papers, Ufa, 1990, pp. 35 - 49.

Semikhatov, M.A., Shurkin, K.A., Aksenov, E.M., *et al.*, New Stratigraphic Scale of the USSR Precambrian, *Izvestiya AN SSSR. Ser. Geol.*, 1991, no. 4, pp. 3 - 16.

Sokolov, B.S., Biochronology and Stratigraphic Boundaries, *Problemy Obshchei i Regional'noi Geologii* (Problems of General and Regional Geology), Nauka, Novosibirsk, 1971, pp. 155 - 178.

Sokolov, B.S., Vendian System: Position in the Stratigraphic Scale, *27 MGK. Stratigrafiya. Sektsiya C.01. Doklady* (27th Intern. Geol. Congress Stratigraphy. Section C.01. Reports), Moscow, 1984, vol. 1, pp. 111 - 127.

Sokolov, B.S., Vendian System. Geological History and Paleontology, *Vendskaya Sistema. Istoriko-Geologicheskoe i Paleontologicheskoe Obosnovanie. T. 2. Stratigrafiya i Geologicheskie Protsessy* (Vendian System. Historico-geological and Paleontological Substantiation. V. 2. Stratigraphy and Geological Processes), Nauka, Moscow, 1985, pp. 199 - 214.

South African Committee for Stratigraphy. Stratigraphy of South Africa. P. 1 (Kent, L.E., Compiler), *Handbook Geol. Surv. S. Africa*, 1990, vol. 8, p. 690.

Stockwell, C.H., Proposal for Time Classification and Correlation of Precambrian Rocks and Events in Canada and Adjacent Areas of the Canadian Shield. P. 1. A Time Classification of Precambrian Rocks and Events, *Geol. Surv. Canada*, Pap. 80 - 12, 1982, p. 135.

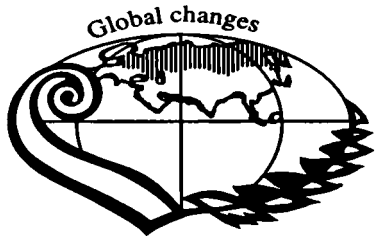
Sun Dazhong and Lu Songnian, A Subdivision of the Precambrian of China, *Precambr. Res.*, 1985, vol. 28, no. 1, pp. 137 - 162.

Trendall, A.F., The "Geological Unit" (G.U.) – A Suggested New Measure of Geologic Time, *Geology*, 1991, vol. 19, no. 3, p. 195.

Walliser, O.H., Toward a More Critical Approach to Bio-Events, *Lecture Notes in Earth Sciences, vol. 8. Global Bio-Events*, Springer, 1986, pp. 5 - 16.

Winter, H., de la R., Kaapvaal Chronostratigraphy Should Be Based on Known Rock Successions Rather than Any Proposed Estimated Time Scale, *S. Afr. J. Geol.*, 1990, vol. 93, no. 516, pp. 864 - 868.

Stratigraphy and Geological Correlation, Vol. 1, No. 1, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 20, No. 2, 1993, pp. 21 - 34.
Original Russian Edition Copyright © 1993 by Stratigrafiya. Geologicheskaya Korrelyatsiya, Chumakov.
English Translation Copyright © 1993 by Interperiodica.



Evolution of the biosphere

Riphean Middle Siberian Glaciohorizon

N. M. Chumakov

Geological Institute, Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 109017 Russia

Manuscript received March 20, 1992

Abstract – Riphean deposits on the western slope of the Aldan Shield and over the periphery of the Baikal fold belt were found to contain glacial deposits with similar stratigraphic positions which can be grouped in the Middle Siberian Glaciohorizon. Its age is Middle Riphean or, less probably, Late Riphean. The Middle Siberian Horizon may be the first evidence of Middle Riphean glaciations on Earth or Upper Riphean glaciations in Northern Asia.

A number of major glaciations that followed one another at the end of the Precambrian may be one of the most prominent, if not adequately understood, phenomena in the geologic record. Glaciations were so typical of the Late Precambrian that the International Subcommission on Precambrian Stratigraphy suggested calling one of its divisions Cryogenian (Cowie *et al.*, 1989). For all the studies done, the number, succession, areas of distribution, and ages of Late Precambrian glaciations need to be specified further. Of particular interest in this respect are the Late Precambrian glacial strata described recently in the Baikal Mountains and the adjacent regions of the Mid-Siberian Plateau (Fig. 1).

RIPHEAN GLACIAL DEPOSITS OF THE REGION

Glacial strata were found in three Riphean[1] formations: Nichatka, Bolshoi Patom, and Dzhemkukan formations. In addition, glacial strata may also be present in the Riphean strata of the Southern Pribaikal'e.

The Nichatka Formation is distributed on the western slope of the Aldan Shield in the southwestern part of the Berezovka Basin where it composes the base of a gently inclined northwestern Riphean to lower Paleozoic sedimentary cover, and lies directly on the crystalline basement. Additionally, some outcrops of the formation extend northwards to the Nichatka Lake along the zone of faults which separate the Berezovka Basin from the Patom fold zone. In this region, the Nichatka Formation unconformably overlies the Middle Riphean Ballaganakh Subgroup; it is folded, deformed by faults and minor thrusts, and in some places with shear cleavage. The top of the formation is placed at the base of a member of typically red, very thinly laminated dolomites and marls of the Kumakhulakh and Barakun formations. This member conformably overlies the Nichatka Formation, serving as a perfect marker to help trace the formation (Fig. 2). The territory hosting the Nichatka Formation is about 70 km from west to east, and over 80 km from south to north.

The sequences of the Nichatka Formation in the eastern cratonic part of this territory and in the western fold-thrust zone are notably different. In the western zone, the sections are relatively uniform while in the east, they are, in contrast, widely variable, reflecting the diversity of the environments there.

The Nichatka Formation on the craton contains three main types of sections: (1) composed of alternating members of massive, sometimes bedded diamicts [2] and laminated rocks; (2) composed of conglomerates and grit sandstones; (3) composed of rhythmically intercalated graded beds of conglomerates, diamicts, sandstones, and siltstones. Considered below are the cratonic types of sections in the order they are listed above.

[1] The Riphean embraces the Mesoproterozoic and most of the Neoproterozoic (except for the Terminal Proterozoic System, or Vendian) of the International Stratigraphic Scale, and is subdivided into three parts: Lower Riphean (1650 ± 50 – 1350 ± 20Ma), Middle Riphean (1350 ± 20 – 1000 ± 50 Ma), and Upper Riphean (1000 ± 50 – 650 ± 20 Ma).

[2] The notion of diamict (Harland *et al.*, 1966) embraces all clastic sediments and rocks, with any degree of lithification, of any texture, structure, or genesis, which are composed of relatively fine-grained groundmass (matrix) and much larger fragments (phenoclasts, stones) dispersed in it. Diamict is almost synonymous to tilloid, but in contrast to the latter, has not been spoilt by incorrect usage (Pettijohn, 1957).

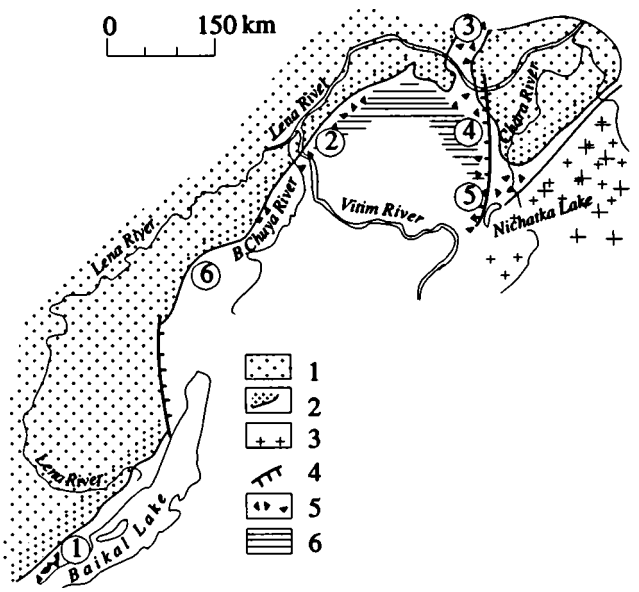


Fig. 1. *Deposits of the Middle-Siberian Glaciohorizon in the Baikal Mountains.*

1 – Paleozoic strata; 2 – Paleozoic-Riphean boundary; 3 – Lower Proterozoic and Archean; 4 – fault zones; 5, 6 – facies of the Mid-Siberian Glaciohorizon: 5 – predominantly diamicts and conglomerates, 6 – predominantly shales.

A characteristic sequence of type (1) is a section in the northwestern shore of the Nichatka Lake. There, members of massive reddish-brown diamicts are 5 to 70 m thick. They contain many stones with abrasive glacial faces showing on the surface variable subparallel lengthwise striae and scars (Figs. 3a and 3b). Some scars on the stones of crystalline rocks form a continuous row of fine spear-shaped glacial chattermarks (Fig. 3c). In some strata of massive diamicts, elongated stones are positioned almost parallel to bedding, and have largely northerly and westerly orientations. Laminated rocks beneath diamicts are sometimes disharmonically folded and brecciated; the structural pattern is complicated by thrusts resembling cryotectonic deformations (Fig. 4c). The diamicts may occasionally contain singular thin lenses of sandstone and almost undeformed rafts of bedded siltstone up to 1.7 m long.

Beds of similar massive diamicts are also present in the Nichatka Formation on the left banks of the Chara River (the lower course of the Mokryi Kumakhulakh). They differ from the above-described in their occasionally grey color and the presence of small erosion channels filled with graded beds of terrigenous sediments. In addition, there are also bedded diamicts there. In contrast to the massive varieties, their badly sorted matrix shows hardly discernible and fairly thin bedding. These diamicts are often linked through gradual transitions (both vertically and laterally) with laminated siltstones and clayey rocks with embedded stones.

Members of laminated red and variegated siltstones and clayey rocks range in thickness from 1.5 to 20 m. They intervene between diamict members and often contain dispersed, faceted and striated dropstones varying in size from a large granule to a boulder and showing typical accompanying structures (Fig. 4b). Very thinly laminated (0.5 - 5 mm) mudstones occasionally contain interbeds of graded fine- and medium-grain sandstones and pebble conglomerates with small fragments of mudstones. In the lower course of the Mokryi Kumakhulakh River, laminated members sometimes contain plano-convex or biconvex diamict lenses 0.2 - 1.5 m thick and from a few meters to 15 m long.

The stones dispersed in diamicts and in laminated silty and clayey rocks are identical in lithology, size, and shape. Their lithology is highly diverse and varies over the area. In the east, along the Chara River and in the middle course of the Sen' River that joins the Nichatka Lake to the Chara River, the stones are dominated by fragments of various gneisses that may be large (up to 3 × 2.5 m across), often with sharp corners and a flaggy shape (Fig. 4a). This suggests that the main provenances were the adjoining parts of the Aldan Shield. This conclusion seems to conform with the predominantly western and northwestern orientation of the elongated stones in diamicts there. Farther west, in the vicinity of the Nichatka Lake, stones, including the largest ones, are represented mostly by various granitoids similar to those composing the Lower Proterozoic basement in that area. Judging by the orientation of elongated fragments in diamicts, the provenance must have been somewhere to the south and southwest. Dolomite stones may sometimes be fairly large, and they are ubiquitous in diamicts and the associated rocks on the left bank of the Chara River, but rare near the Nichatka Lake. Dolomites are represented by oncolitic and stromatolitic varieties, with no known sources around, so for this particular area, such stones may be classified as erratics. The association of massive diamicts and bedded rocks with dropstones that occurs almost horizontally and is widespread, and the abundance of large, erratic, faceted and striated stones therein, consistently indicate the glacial origin of these rocks. A unimodal orientation of elongated stones in some massive diamicts, and cryotectonic deformations at their base are evidence of basal tillites strata in the section of the Nichatka Formation. The fairly gradual transition between laminated rocks and the overlying massive diamicts is supported in other cases by the presence in laminated rocks of rare to abundant dropstones and individual lenses of diamicts (iceberg tills?). Laminated rocks with dispersed stones give way, in turn, to bedded diamicts and then to massive diamicts, or directly to massive diamicts. As already noted, the latter may occasionally show shallow erosion channels filled in by thin graded deposits with dispersed pebbles. Such sequences can be interpreted as the results of the appearance of icebergs in the basin, with their number growing, and later-of the progradation of the ice shelf into the basin and accumulation underneath of various

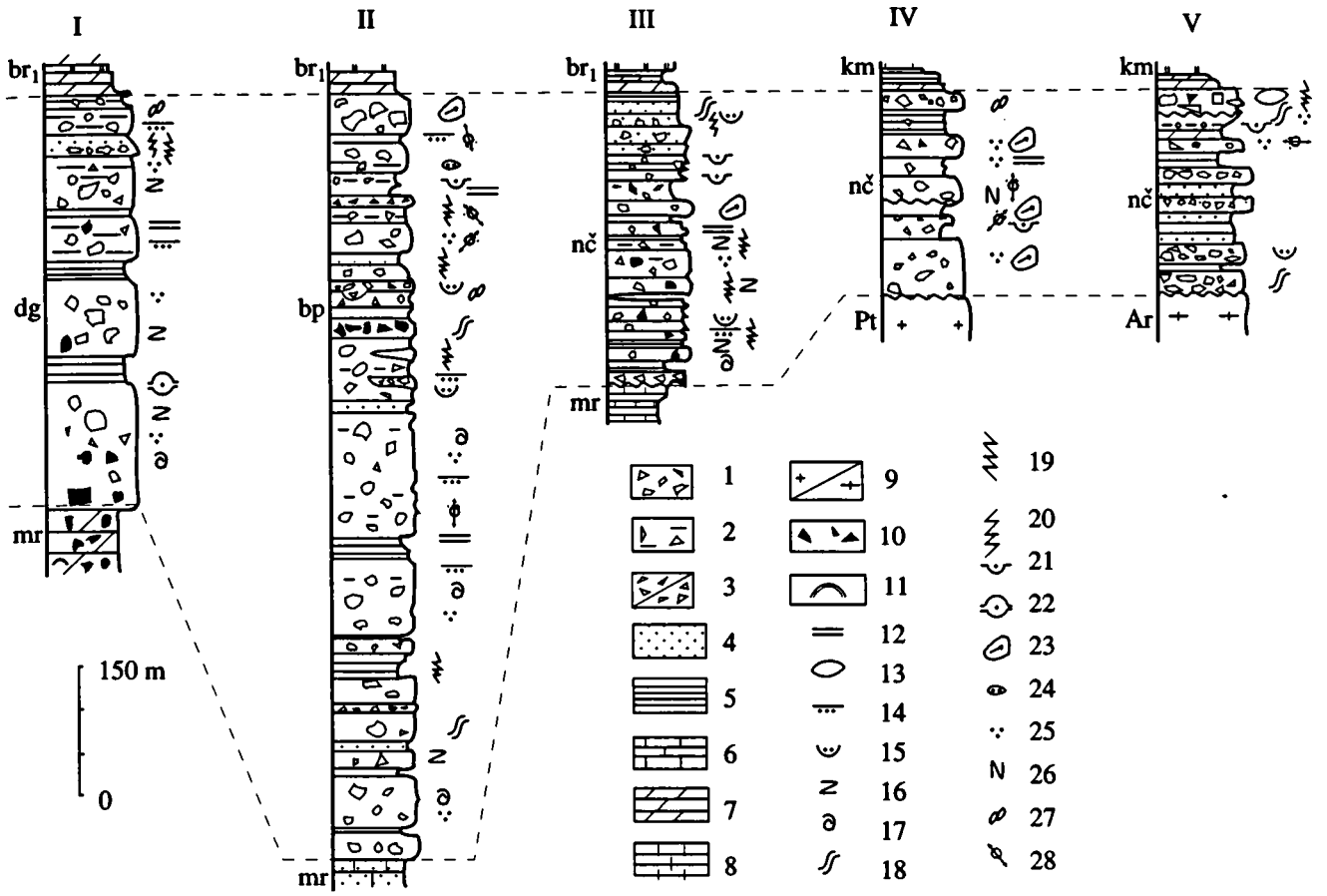


Fig. 2. *Key sections of the Middle-Siberian Glaciohorizon in the Northern Baikal and Patom highlands.*
I – Bol'shaya Chuya River; II – Bol'shoi Patom River (Ura Rise); III – Oporny spring, upper stream of the Dzhelinda River (eastern slope of the Longdor Rise; IV – Nichatka Lake (SW Berezovka Basin); V – Chara and Mokryi Kumakhulakh rivers (southern Berezovka Basin). Formations: dg – Dzhemkukan; bp – Bolshoi Patom; nc – Nichatka; mr – Mariinskaya; br – Barakun; km – Kumakhulakh. 1 – massive diamicts; 2 – bedded diamicts; 3 – conglomerates and breccias; 4 – sandstones; 5 – shales and siltstones; 6 – limestones; 7 – dolomites; 8 – marls; 9 – granites and gneisses; 10 – stones of dolomites; 11 – stromatolites; 12 – lamination; 13 – lenses; 14 – lenses with a gradual lower contact; 15 – erosion cuts; 16 – submarine slump structures and deposits; 17 – globes and other fragments of sedimentary beds of a submarine-slump origin; 18 – cross-bedding; 19 – normal rhythms; 20 – inverse rhythms; 21 – dropstones; 22 – singular stones; 23 – striated stones; 24 – till pellets; 25 – stone nestles; 26 – cryotectonic textures; 27 – split stones; 28 – predominant stone fabric.

aquatills: melt-out or waterlain tills (Dreimanis, 1979; Gravenor *et al.*, 1984), and submarine basal tills (Link and Gostin, 1981). Turbidity or other density flows arose on the slopes of aquatillites accumulations and cut in channels which were to be subsequently filled by the deposits of minor debris and turbidity flows. Occasionally anchored, the ice shelf produced the aforementioned cryotectonic deformations at the base of some of the diamicts. As a result, the association of massive diamicts and laminated rocks with dispersed stones can be viewed as an alternation of distal and proximal glacio-basinal facies.

The presence of proximal glacial facies in the Nichatka Formation is proved also by the lateral relations between diamicts and conglomerates, typical of type (2) of the sections. Occasionally, members of massive diamicts and laminated siltstones can be seen rapidly wedging out, or in part, replaced by thick conglomerates (Fig. 5). In the case illustrated in Fig. 5, the lower wedging members of massive diamicts and laminated siltstones must have been deposited in buried valleys (fjords?) or other similar depressions of the ancient topography, while the upper diamict member must have been lain between submarine fans produced by subglacial streams of outlet glaciers or the ice shelf. The conclusion about the submarine fans is based on the fact that conglomerates which replace diamicts along the strike and grade upsection in sandstones compose large lens-shaped bodies, 200 to 350 m thick and a few kilometers across. The conglomerates show large boulders and a poorly sorted, but abundant matrix, and contain scattered lenses and beds of normal fine- and medium-boulder conglomerates and cross-bedded sandstones. The overlying thick-bedded gravel arkosic sandstones comprise numerous lenses of narrowly intercalated, poorly sorted conglobreccias and gravel

Table 1. Straigraphic position of Riphean dolomites over the periphery of the Baikal Mountains (data by the author, Bozhko (1976); Golovenok *et al.* (1963); Zhuravleva *et al.* (1959); Keller *et al.* (1967), Khomentovskii *et al.* (1972), Dol'nik *et al.* (1987, 1990))

Age	Southern Peri-Baikal	Northern Baikal Highland (Bol'shaya Chuya and Vitim rivers)		Ura Uplift (Lena and Bol'shoi Patom rivers)	Berezovka Basin	
					Dzhelinda River	Nichatka Lake – Chara River;
ϵ_1	Bel'sk Fm., *Elganellus, Malykania* limestones, dolomites	Nokhtuysk Fm., IVs, dolomites, marls, 600 m		El'gyan Fm., *Malykania* limestones 50 m		El'gyan Fm., *Elganellus, Malykania* limestones, 50 m
	Usol'e Fm., dolomites, breccias, 300 m			Macha Fm., dolomites, 160 m		Yuedey Fm., dolomites, varicolored marls, 300 m
Vendian (V)	Ayankan Fm., IVs, dolomites, shales, 150 m			Nokhtuysk Fm., IVs, dolomites, marls, 300 m		
	Kurtun Fm., IV m, M, sandstones, siltstones, 250 m	Tinnaya Fm., IV s, m, limestones, dolomites, shales, 250 m		Tinnaya Fm., IV s, m, dolomites, limestones, breccias, 500 m	Tinnaya Fm., > 200 m, dolomites, breccias, limestones	Tinnaya Fm., IV s, dolomites, breccias, limestones, 250 m
	Ushakovka Fm., 609 G, sandstones, conglomerates, shales, 1300 m	Zherba Fm., quartz sandstones, 350 m;		Zherba Fm., 590 G, quartz sandstones, 300 m	Zherba Fm., quartz sandstones, 200 m	Zherba Fm., quartz sandstones, 50 m
	IV m					
Upper Riphean (R_3)	Baikal Group: Kochergat Fm., shales, sandstones, III, III m, 1000 m	Vendian (V)	Zhuya Group: Chencha Fm., III s, light limestones, siltstones, 600 m	Chencha Fm., III s, pale stromatolitic, oolitic limestones,	Chencha Fm., III, pale stromatolitic, oolitic limestones, 500 m	Torgo Fm., III s, m, pale stromatolitic, oolitic, limestones, red marls at the base, 250 m
			Nikol'skaya Fm., III, red siltstones, marls, 500 m	Nikol'skaya Fm., III, red and green marls, 350 m	Nikol'skaya Fm., III, red marls, 150 m	
Middle Riphean (R_2)	Uluntuy Fm., dolomites, limestones, shales, III s, III m, 500 m		Dal'naya Taiga Group: Valukhta Fm., II s, shales, sandstones, limestones above, 500 m	Valukhta Fm., II s, shales, and dolomites, limestones above, 1000 m	Sen' Fm., II s, dolomites, limestones, 400 m	Sen' Fm., II s, dolomites, limestones, 350 m
	Goloustenaya Fm., dolomite, sandstones, shales, 400 m		Barakun Fm., II s, dolomites, limestones, shales, at the base sandstones, marls, cream dolomite, 600 m	Barakun Fm., II, black limestones, shales, at the base sandstone and cream dolomite, 900 m	Barakun Fm., II, black limestones, shales, at the base sandstones, red marls and dolomites, 500 m	Kumakhulakh Fm., II, black limestones, siltstones, at the base red siltstones and dolomite
	Diamicts, shales, sandstones, 100 m		Dzhemkukan Fm., diamicts, sandstones, conglomerates, 500 m	Bol'shoi Patom Fm., diamicts, sandstones, conglomerates, 1000 m	Nichatka Fm., diamicts, sandstones, conglomerates, 400 m	Nichatka Fm., diamicts, sandstones, conglomerates, 300 m
			Ballaganakh Group: Mariinskaya Fm., II s, dolomites, breccias, 150 m	Mariinskaya Fm., II, limestones, calcareous sandstones, 500 m	Mariinsk Fm., II s, black shales, limestones, subordinate dolomites, 100 m	
			Bugarikhta Fm., sandstones, shales, 300 m	Bugarikhta Fm., sandstones, gravelstones, shales, >1000 m	Sandstones, conglomerates 500 m	
			Khaiverga Fm., shales, sandstones, 400 m			
			Kharlukhtakh Fm., conglomerates, sandstones, diamicts above, 1300 m			
			Medvezhevka Fm., amygdaloid metadiabases, 700 m		Medvezhevka Fm., sandstones, diamicts, 250 m	
R_{2-1}			Purpul Fm., quartzites, schists, 550 m		Purpul Fm., quartzites sandstones, schists, 1500 m	Purpul Fm., red quartzites, 50 m
R_1			Teptorgo Group: Chuya granites		Kodar granites 1790 γ	Nichatka granites 1780 γ; gneisses, granites
PR_1 + Ar	Granites					

Notes: II, III, IV, V – microphytolites of the Middle, Upper Riphean, Vendian and Cambrian, respectively; II s, III s, IV s, V s – same, plus the respective stromatolites; III m, IV m – microfossils of the Upper Riphean and Vendian, respectively; IV M – nonskeletal Metazoa; 590 G – K/Ar age on glauconite; 1790 γ – Pb/Pb age on granites.

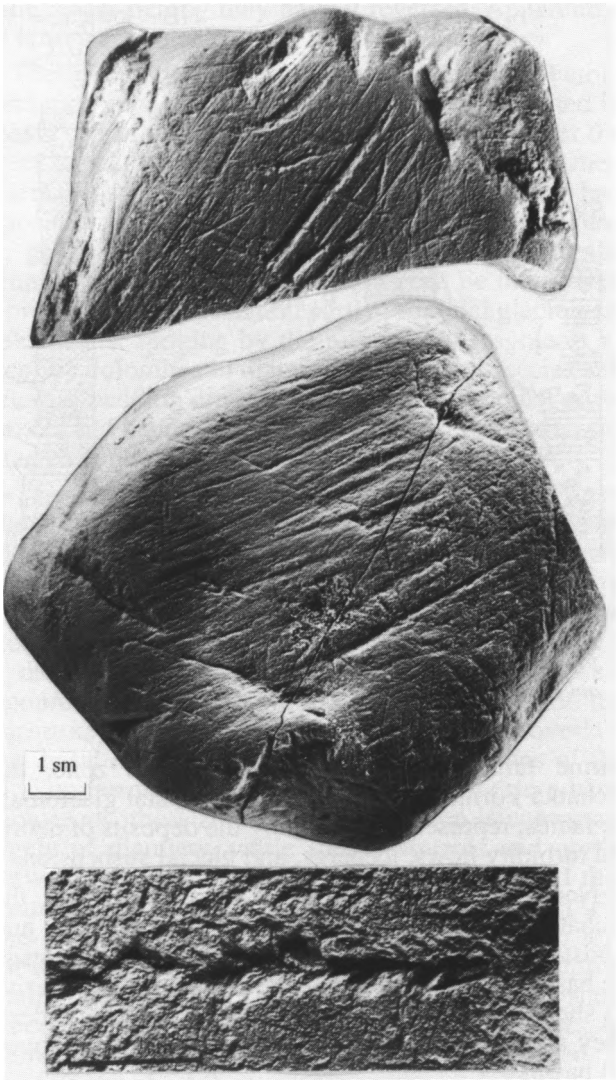


Fig. 3. *Striation and marks on faceted stones of the Nichatka Formation, near the Nichatka Lake, north of the Shirik Creek. The lower image shows spear-shaped glacial chattermarks on a granite pebble.*

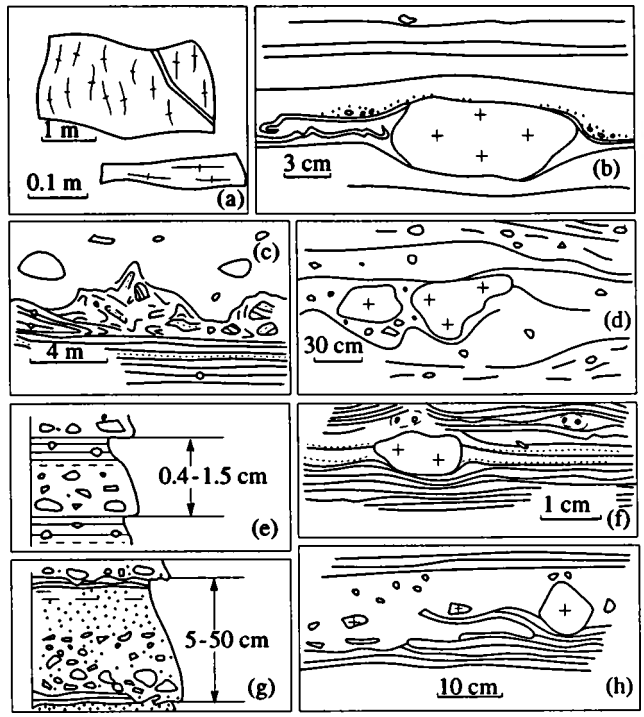


Fig. 4. *Some typical boulders and structures of glacial and glaciomarine deposits of the Nichatka and Dzhemkukan formations.*

a – large and angular pieces of gneisses from the Nichatka Formation, Mokryi Kumukhulakh River, b – a dropstone in laminated siltstones, same formation and locality, c – cryotectonic deformations at the base of one of massive diamict beds of the Nichatka Formation, near the Nichatka Lake, north of the Shirik Creek, d – a nestle of boulders in sandy diamicts with traces of pressure below and draping above (dropstones), the Bol'shoi Patom Formation, Ura River, 5 km downstream the Iligir' River, e – a structure of a typical rhythm unit in the middle of the Nichatka Formation, Dzhelinda River, Oporny Spring, f – a dropstone and till pellets in laminated siltstones of the Bol'shoi Patom Formation, estuary of the Bol'shoi Patom River, g – a structure of a typical rhythm unit in the Bol'shoi Patom Formation, estuary of the Bol'shoi Patom River, h – a rip structure in siltstones at the base of the graded diamicts, Bol'shoi Patom Formation, estuary of the Bol'shoi Patom River.

sandstones. These deposits fill in flat and broad erosion channels. Within the 65 km stretch from the Nichatka's western shore to the Chara's eastern bank, three such fans were identified. Members of massive diamicts may sometimes wedge into the edges of these fans (Fig. 5, etc.); or, occasionally, overlap them (Chara River).

Type (3) of Nichatka sections on the craton consists mostly of rhythmic, graded beds of red clastic rocks. Their composition and structure are diverse, ranging from diamicts to siltstones, although coarse clastics seem to predominate. This section type is typical of a small area in the mid-stream of the Sen' River. The lower part of the formation is marked there by a normal rhythmic alternation of 0.5 - 2 m diamicts and laminated siltstones with dispersed stones. In the middle part of the formation, the rhythms are composed (upward) of diamicts, gravelstones and sandstones, while in the upper part, there are large-boulder diamicts, boulder and pebble conglobreccias, fine- to medium-pebble conglomerates, gravelstones (rarely sandstones) and siltstones. These rhythms vary in thickness from 2 to 6 m. The bulk of rhythms consists of diamicts which contain blocks of gneisses 2.5 × >3 m across, and conglobreccias abundant in randomly oriented, poorly rounded and unrounded platy fragments of crystalline rocks up to 1 m long. Additionally, the conglobreccias may contain large (up to 1.2 × 0.4 m) curved rafts of underlying diamicts. The rhythms and their basal layers, in particular, may be lens-shaped as they fill shallow erosion depressions in the underly-

ing rocks. The upper fine-grained layers of the rhythms drape minor features and bulging boulders at the top of diamicts and conglobreccias, levelling off the surfaces of erosion cuts. The siltstones that crown the rhythms sometimes contain typical dropstones, occasionally striated. The general impression is that the middle and upper parts of these types of sections are, in fact, a sequence of inset erosion channels filled by the deposits of debris flows and ultimately levelled off by suspension sediments into which stones melt out of the ice shelf or icebergs fall. The coarse-clastic composition of the deposits, as well as their local extension, the abundance of erosion channels and the presence in conglobreccias of deformed intraformational rafts suggest that these types of sections formed in a submarine canyon apparently associated with one of the fans.

In the western fold-thrust zone, the Nichatka Formation is composed mostly of rhythmically alternating graded beds of boulder diamicts and laminated siltstones or shales containing dispersed stones. Massive diamicts are subsidiary in the lower and middle parts of the formation there.

The thickness of the rhythms in the western sections varies commonly from a few centimeters to 2 - 3 m (Fig. 4e). The lower boundary of the diamict beds that compose the base of the rhythms is usually abrupt, sometimes with traces of erosion. The lower part of the diamicts is often enriched in fragments of underlying rocks. In some cases, diamicts form lenses and fill in shallow and broad erosion channels. Laminated siltstones and shales at the top of the rhythms show in turn a graded bedding of a smaller magnitude (from a few millimeters to a few centimeters). At the base of these second-order rhythms is a very thin layer of sandstones. The stones dispersed in laminated rocks form numerous nests, and contain typical dropstones and occasionally striated boulders. These features suggest that intercalated diamicts and laminated rocks in the western sections can be interpreted as an alternation of deposits formed by debris and turbidity flows, which contained the dispersed material of iceberg rafting. The aforementioned subsidiary members of massive diamicts in the lower and middle parts of the formation must be accumulations of glacial suspension and iceberg coarse clastics. Attesting to their glacial-basinal origin are the dispersed and striated stones, as well as traces of the matrix's reworking by weak currents and submarine slumps.

Layers of sandstones with inverse graded bedding, which occur in the upper part of the western Nichatka sections, seem to be deposits of grain flows.

In summary, the above facts suggest that the Nichatka Formation is represented by glacial deposits accumulated in a basin. It is very likely that the basin was marine since the Nichatka Formation is both under- and overlain by thick carbonate formations (Mariinskaya, and Kumakhulakh or Barakun formations respectively). In the eastern platform zone, the Nichatka Formation is composed mostly of proximal glaciomarine facies, fairly coarse-clastic featuring sub-

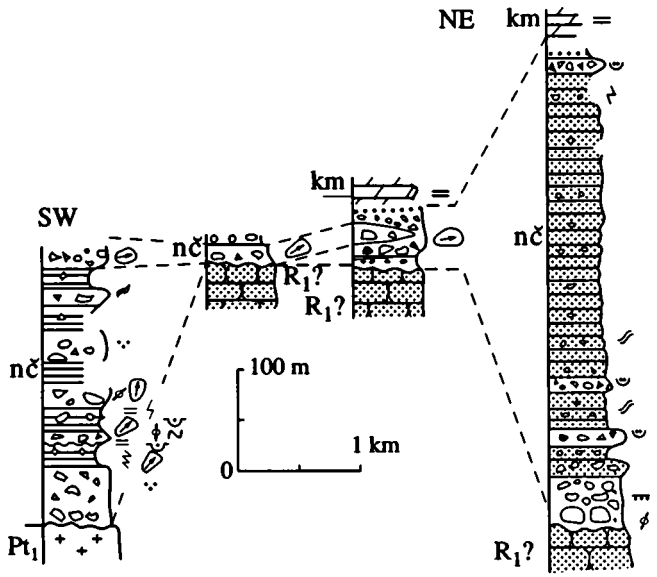


Fig. 5. *Stratigraphic and facies changes in the Nichatka Formation along the Nichatka and Dekanda lakes divide. For legend see Fig. 2. R_1 – brown-red quartzites of the Puroul Formation.*

marine fans. In the western fold-thrust zone, the Nichatka Formation is dominated by distal glaciomarine facies, represented largely by the deposits of debris and turbidity flows, icebergs, and glacial suspension.

North, at the edge of the Patom Fold Zone, the Nichatka deposits are replaced by black shales and sandstones of the Dzhemkukan Formation. Its shales in the basins of the Bul'bukhta and Bol'shoi Barakun rivers show occasional horizons with dispersed boulders. They are probably signs of iceberg rafting in an open sea basin.

The Bol'shoi Patom Formation is exposed at the cores of two anticlines of the Ura Uplift along the Lena River and its tributaries (Fig. 1). The formation composed largely of diamicts is about 1000 m thick; it rests on sandy limestones of the Mariinskaya Formation and is covered conformably by Barakun dolomites. Since the lithology and genesis of the Bol'shoi Patom deposits have been discussed recently in a separate paper (Chumakov and Krasil'nikov, 1991), it may be sufficient to present only the conclusions on the formation environments, and the facts that led to the conclusions.

The formation is dominated by massive diamicts, locally with striated boulders or signs of erosion by bottom current and submarine slumps. Common are bedded diamicts, and siltstones with dropstones and till pellets (Fig. 4d and 4f). Quite common are members of rhythmically alternating graded beds of diamicts and sandstones definitely deposited by debris flows (Fig. 4g and 4h). Locally, there are boulder conglomerates filling erosion channels, beds of conglobreccias and members of thin-grain sandstones with a cross-bedding of the ripple-drift type. These sandstones form large lenses with a fan-shaped cross-bedding inside,

while lenses nearby may have it reversed. Apparently, the lenses are outer parts of submarine fans.

The associations and character of Bol'shoi Patom Formation suggest that the formation was deposited in a basin (presumably, marine and fairly deep) under the direct impact of icebergs or shelf ice, and glacial melt waters. Most of the sediments that the above agents had deposited, were reworked on submarine slopes by clastic, grain, and turbidity flows, and bottom currents. Generally, Bol'shoi Patom deposits can be interpreted as predominantly distal and partly proximal glaciomarine deposits. Judging by the presence of fragments of oncolitic dolomites in diamicts and conglobreccias and the cross-bedding dip in sandstones, the principal provenance, and hence, glaciation areas must have been situated east, north- and southeast of the Ura Uplift.

The northwestern margin of the Baikal-Patom Highland. Glacial strata at the given stratigraphic level belong here to the Dzhemkukan Formation. They were traced between the upper course of the Malaya Chuya River in the southwest and the Lena-Bol'shoi Patom interfluve in the northeast for at least 250 km (Fig. 1). In this region, the Dzhemkukan Formation occurs unconformably on dolomites and limestones of the Mariinskaya Formation, and is conformably overlain by a basal member of dolomites or limestones of the Barakun Formation. In the extreme northwestern band of outcrops, the Dzhemkukan Formation is composed largely of diamicts, while southeastwards and northeastwards they are replaced by black shales, and thin laminated distal turbidites and sandstones. Diamict sections of the Dzhemkukan Formation may be up to several hundreds of meters thick (Fig. 2). They comprise mostly massive diamicts, locally with indistinct signs of bedding, lenses of conglomerates and sandstones, or submarine slump patterns. The stones dispersed in diamicts are 1 - 1.2 m across, composed of various rocks, mostly plagioclase granites, Mariinskaya dolomites, quartzites, gneisses, etc. The stones are variably rounded, although on average, they are poorly rounded or semi-rounded; there are also slightly rounded and split stones. The stones often occur in small groups (nests). In the diamicts there are rare layers and beds of shales also with embedded stones (Krasil'nikov, 1990). In general, Dzhemkukan diamicts strongly resemble Bol'shoi Patom diamicts, the only difference being fewer sandy grains in the matrix. Along with the identical stratigraphic position, similar lithology implies that the Dzhemkukan Formation in the northwestern margin of the Baikal-Patom Highland must have also been deposited in a glaciomarine environment.

Southern Pribaikal'e. In the margin of the Baikal-Patom Highland, between the upper stream of the Malaya Chuya River and the Southern Pribaikal'e zone, the Riphean sequence starts with a carbonate-terrigenous Goloustnaya Formation, which, as it will be shown below, is primarily a stratigraphic analogue of the Barakun Formation. The Goloustnaya Formation unconformably overlies the Lower Proterozoic rocks. Strata contemporaneous with those of the Dzhemkukan Formation did not accumulate here, or else were eroded. Only in the southern part of the western Pribaikal'e zone, in the north of the Bugul'deika Depression, does a sequence of sedimentary and volcanic rocks occur again between the Goloustnaya Formation and the Lower Proterozoic granites; the sequence is composed of diamicts, shales, gravel conglomerates, and diabase porphyrites. Massive diamicts form 100 m thick members in the lower part of the section. The lithology of the diamicts is typical. They consist of a fine-grain chlorite–sericite–siltstone matrix which contains scattered angular and variably rounded stones of various granites, sometimes up to 1.3 m across. The stones are dominated by the varieties of granites that underlie the diamicts. There are occasional members of variegated shales with fine dispersed fragments of granites, and poorly sorted fine-pebble conglomerates which resemble the diamicts' matrix. The section of the pre-Goloustnaya rocks is crowned by metamorphosed diabase porphyrites. The thickness of the described sequence was formerly assumed as inconsiderable, and the relevant rocks were interpreted as forming the basal portion of the Goloustnaya Formation (Salop, 1964), or else included in the Middle Proterozoic Khoty Formation. It was shown, however, that the pre-Goloustnaya sequence is several hundreds of meters thick and apparently covered by Goloustnaya carbonates (Bozhko, 1976).

This fact, as well as the presence of diamicts in the section, suggests that a part of the pre-Goloustnaya deposits in the Bugul'deika Depression must be a stratigraphic and facies counterpart of the Dzhemkukan Formation of the Patom Highland.

Correlation of glacial deposits. The above-described glaciomarine strata of the Patom and Northern Baikal highlands are part of the Riphean Patom Supergroup. Due to similar sections, the presence of a number of typical markers, fossils, and transitional sequences, the groups and formations of the supergroup are easily traced (and by geological survey as well) over the periphery of these highlands. They were traced from the Nichatka Lake in the east as far as the Bol'shaya Chuya basin in the west (Table 1). The discrepancies between the correlations suggested by various authors pertain solely to the boundary portions of the formations, and to their ranges. One of the marker intervals in the supergroup's section is the Zhuya Group whose units, although varying in thickness, consistently retain their lithology unchanged throughout the peripheral segment of the highlands. Another marker interval is the Barakun and Kumakhulak formations. True, their lower boundary in the sections studied is not unanimously regarded. In the Bol'shaya Chuya Basin, the lower terrigenous-carbonate member of the formation is sometimes regarded as an upper member of the Dzhemkukan Formation (Belogur and Tikhonov, 1970). However, this assumption neglects the fact that in the stratotype section of the Bol'shoi Barakun River and the correlative sections of the Zhuya River, Ura Uplift and Dzhelinda River, the Barakun Formation has

a three-member structure, composed of a lower calcareous-dolomite member, a middle sandstone-shale one, and an upper calcareous one. Some authors treated these beds as members (Chumakov, 1959), while others assigned them the status of formations (Bobrov, 1979). In the margins of the Patom Fold Zone (Ura, Dzhelinda, and Bol'shaya Chuya rivers), the thickness of the Barakun Formation and particularly of its counterpart, the Kumakhulakh Formation, is reduced, mostly at the expense of their lower members; the proportion of sandstones in the middle member is commonly higher (Bol'shaya Chuya, Ura, Bagayukhta, and Sen' rivers), while the rocks sometimes look slightly varicolored (Bol'shaya Chuya, Bagayukhta, and Sen' rivers). The lower member of the Barakun Formation is notably reduced in thickness (down to a few dozens of meters); limestones there are partly (Lena, Bol'shoi Patom, Ura, and Turukshtakh rivers) or fully replaced by shales, siltstones, dolomitic marls and dolomites (Bol'shaya Chuya, Dzhelinda, Bagayukhta, Sen', Mokryi Kumakhulakh rivers, and Nichatka Lake). In most sections of the Berezovskaya Basin, rocks of the lower member are red or varicolored. Very typical is a pile of dolomites at the base of the lower member of the Barakun and Kumakhulakh Formations, directly on the diamicts of the Dzhemkukan, Bol'shoi Patom and Nichatka formations. This basal pile is commonly composed of laminated, pink or cream-colored aphanic or pelitomorphic calcareous dolomites. They may sometimes show fine intrastratal carinate folds (tepee structures). This pile is commonly about several meters to a few dozens of meters thick; however, due to its typical pattern and persistence, it is often a good marker for mapping and correlation. In terms of lithology and position, this pile is a typical cap dolomite which crowns many upper Precambrian glaciomarine formations of the Vendian and Riphean (Chumakov, 1978, 1990; Hambrey *et al.*, 1981). The cap dolomites must have originated at the time of fast-rate warming and the associated deglaciations and post-glacial transgressions ("terminations" for Quaternary glaciogeologists). Geologically, instantaneous glacio-eustatic sea-level rise and inundation of the coasts may account for a wider distribution of many cap dolomites compared to the underlying glacial formations. For instance, cream-colored bedded dolomites of the Nuccaleena Formation, i.e., southern Australia, overlying Marinoan glacial strata and being on average 5 - 20 m thick, spread over an area more than 600 × 600 km. They stretch from east to west from the Barrier Ridge over the basement of the Curnamona median massif across the entire Adelaide miogeosyncline as far as the western cratonic framing of the latter, the so-called Stuart Shelf. At the shelf edge, the Nuccaleena Formation overlaps the thin glacial strata, a variety of older horizons, and even the crystalline basement (Preiss, 1987). The cap dolomites, that mark the onset of a glacio-eustatic transgression, usually grade upsection, as is the case of the Barakun and Kumakhulakh formations, into variegated marls and siltstones, and then sandstones. This regressive succession seems to reflect the glacio-isostatic rising of the land, which is known to lag behind due to a considerable viscosity of the asthenosphere.[3] Thus, cap dolomites appear to reflect very brief climatic events which, judging by the Pleistocene glaciations, did not last more than a few thousand years. Therefore, Precambrian cap dolomites can be considered as virtually synchronous stratigraphic levels. It follows from the above that in our case, it may be reasonable to assume that the base of the cap dolomites should be the lower boundary of the Barakun and Kumakhulakh formations, both in terms of the regional correlation and in terms of the event stratigraphy.

The next older regional marker is the sandy-calcareous Mariinskaya Formation, which, although replaced by dolomites in the margins of the Patom Fold Zone, still remains a sole carbonate formation in the terrigenous sequence of the lower part of the Patom Supergroup.

A few typical markers are also present in the region below this supergroup (Table 1). Note also, that similarity there is not only between individual formations or members of the Patom Supergroup, but also in their general sequence in sections and in the sequence of their fossils. Thus, Dzhemkukan, Bol'shoi Patom and Nichatka glacial deposits are enclosed between correlative stratigraphic units can, thus, be assumed as contemporaneous. It is suggested that the set of these three glacial formations be combined in a regional climatostratigraphic unit: a glacial horizon (or, for short, glaciohorizon), as it is in the world's many Precambrian deposits (Zubtsov, 1972; Chumakov, 1971, 1978). Naming the new glaciohorizon Middle-Siberian was also suggested (Chumakov, 1991).

The correlation of the Patom Supergroup with the Baikal Group of the Pribaikal'e is not as unambiguous as the correlation inside the Patom Supergroup. There are particularly many discrepancies pertaining to the lower part of the Baikal Group. There are three main points of view: according to the first one, the Goloustnaya Formation corresponds to the entire or the lower part of the Barakun Formation (Golovenok *et al.*, 1963; Keller *et al.*, 1967; Mats and Bukharov, 1967); according to another point of view, the Goloustnaya Formation corresponds to the Dzhemkukan Formation (Salop, 1964), while the third view holds that the Goloustnaya Formation correlates with the Dzhemkukan and Mariinskaya formations (Dol'nik *et al.*, 1987).[4] The first of the listed opinions was substantiated by detailed studies carried out by B.D. Belogur and V.L. Tikhonov in the basin of the Malaya Chuya River. The authors, however, included the two lower members of the Barakun Formation in the upper Dzhemkukan member and

[3] This lag is used in particular to calculate the viscosity of the asthenosphere (Artyushkov, 1967).

[4] Not included here is still another point of view, suggesting a correlation between the Goloustnaya Formation and the upper part of the Valyukhta Formation (Khomentovskii *et al.*, 1972), as the authors have given it up (Khomentovskii, 1990), and there seemed to have been no other supporters of that correlation.

made a conclusion about the correlation between the Goloustnaya Formation and the upper member of the Dzhemkukan Formation. Comparing layer by layer, sections of the Dzhemkukan Formation in the Malaya Chuya basin, Belogur and Tikhonov proved that the dolomite capping the Dzhemkukan conglomerates and diamicts in Bol'shaya Chuya sections corresponds to the basal dolomite of the Goloustnaya Formation at the Chaya River (Belogur and Tikhonov, 1970); hence, the Goloustnaya Formation is postglacial.

The correlation of the Goloustnaya Formation with the postglacial part of the Patom Supergroup is feasible also for general paleogeographical reasons. In the areas between the upper streams of the Malaya Chuya and the Chaya rivers, outcrops of the Patom Supergroup have an almost east-west strike and cross northeast-trending Riphean structural and facies zones. Gradual thinning and ultimate disappearance, of lower Patom beds west of the Malaya Chuya River, seem to reflect an approach towards the coastal zone of the Riphean Patom basin and the source areas. The same is suggested by stoneorientation in diamicts, their coarser grade southwestward, and the direction of cross-bedding in rocks of Dal'naya Taiga Group. Then, facies changes in the upper stream of the Malaya Chuya and in the lower stream of the Chaya River seem to be identical to those recorded on the opposite, eastern side of the Patom Depression, in the Berezovka Basin: there, the thickness of Patom Supergroup decreases eastwards, its lower horizons successively wedge out, while black limestones and shales are gradually replaced by shallow-water dolomites, red marls, and sandstones. A position nearer to the source area is inferred in the glacial deposits from a replacement of distal glaciomarine facies of the Nichatka Formation by its proximal facies, from the orientation of cross-bedding in sandstones and fragments in diamicts and conglomerates, and then from the appearance of fluvial facies and the wedging out of this formation in the right bank of the Chara River.

In the light of these general facts, it is highly unlikely that the glaciomarine Dzhemkukan Formation in the western marginal part of the basin could be partly or fully replaced coastward by shallow-water sandstones and dolomites of the Goloustnaya Formation, as Salop (1964) assumed, and Dol'nik *et al.* (1987) argued. It is much more likely that the strata synchronous with the Dzhemkukan Formation did not accumulate at all in the area of the lower course of the Chuya River and southwards; even if they did, it was under continental conditions, and the thin deposits were denudated or partly reworked by the postglacial transgression into the boulder horizon that is recorded at the base of the Goloustnaya Formation on the Chaya River and to the south of it.

Alongside the aforementioned correlations by Belogur and Tikhonov, these general facts give every reason to accept that the lower part of the Goloustnaya Formation was deposited in a postglacial basin which inundated the peneplanated edge of the craton as a result of the glacio-eustatic sea-level rise. It can be reliably assumed therefore, that the diamicts beneath the Goloustnaya Formation in the Bugul'deika River basin also correspond to the Dzhemkukan Formation. In this case, the diabase metaporphyrites, registered by Bozhko (1976), may prove, despite the different composition, to be close in age to the tuffogenic sequence described by Belogur under the Dzhemkukan Formation in the Malaya Chuya basin.

AGE

The Patom Supergroup and the Baikal Group occur considerably lower than the Lower Cambrian strata with trilobites, and are conformably overlain by Vendian deposits which contain the respective assemblages of microfossils, stromatolites, microphytolites and, locally Pribaikal'e casts of soft-bodied *Metazoa* (Table 1). In turn, the Patom Supergroup is separated by a thick and complexly structured Teptorgo Group, as well as by three hiatuses and unconformities from the Lower Proterozoic granites dated with the U-Pb isotope method on zircons as 1780 ± 30 Ma. There is, thus, little doubt of the Riphean age of the Patom Supergroup. The question is what part of the Riphean the Dal'naya Taiga Group, which comprises glacial strata, corresponds to. Using stromatolite and microphytolites assemblages, most of the researchers have traditionally referred the Dal'naya Taiga Group to the Middle Riphean (Golovenok *et al.*, 1963; Dol'nik and Vorontsova, 1974; Dol'nik *et al.*, 1987; Keller *et al.*, 1967; Stanevich *et al.*, 1990; Chumakov and Semikhatov, 1981). In the opinion of some other authors springing from facts of historical geology and supported by inter-regional geologic correlations through a number of structural and facies zones, and the revision of the stratigraphic significance of many phytolite forms, the Dal'naya Taiga Group, together with the Zhuya Group, refers to the upper Upper Riphean (Khomentovskii, 1990). Finally, a third point of view, based on correlations by means of several species of microfossils, holds that the Dal'naya Taiga Group belongs to the lower Upper Riphean (Stanevich *et al.*, 1990). The isotope Pb-Pb datings of Dal'naya Taiga (Kadalikan Group) metaschists, which yielded the age of 1414 ± 39 Ma are of little help (Sharov *et al.*, 1972). The authors of these determinations treat the value as the age of metamorphism and, consequently, as the minimum limit of the age of sedimentation. However, the real geological significance if any of this dating is doubtful, due to the problem of inherited isotopes from the source rocks.

Hence, results of different stratigraphic methods, which may challenge the validity of the traditional point of view, either fail to match each other, or are unreliable. It appears that the assumption of the Middle Riphean age of the Dal'naya Taiga Group, and above all, of the Middle Riphean age of the Middle-Siberian Glaciohorizon, seem to fit better than others the stratigraphic position of this horizon and the group, which is in the middle of an intricately built and very thick mio-

geosynclinal sequence that occurs between the Vendian and the Teptorgo Group commonly correlated with the Lower Riphean (Golovenok *et al.*, 1963; Dol'nik *et al.*, 1987; Keller *et al.*, 1967; Khomentovskii, 1990).

PALEOGEOGRAPHY

Glacial strata of the Middle-Siberian Glaciohorizon crop out in three isolated regions of the Patom and Northern Baikal highlands (Fig. 1). Structurally, these regions are uplifted marginal portions of the Riphean Patom Depression and those few areas that show marginal facies of the lower formations of the Dal'naya Taiga Group. Over the rest of the territory of the two highlands, these marginal facies were either concealed under younger strata, or highlifted and eroded. Originally, glacial strata must have framed the Patom basin in an almost continuous belt, composed as the above descriptions of the Nichatka and Bol'shoi Patom formations suggest, predominantly by fans of outlet glaciers.

In all three regions of the development of the glaciohorizon's glacial strata, the paleogeography is similar. The composition of pebbles and boulders, variations in their size, orientation of, cross-bedding, stratigraphic and facies variations suggest that the principal source areas were at the time of glaciation on the adjoining blocks of the Siberian Platform which bordered on the Patom Depression on the northwest and east. The evidence has been cited while describing the glacial strata and the Patom Supergroup–Baikal Group correlation.

Toward the center of the Patom Depression, the proximal glaciomarine facies are replaced by distal glaciomarine, and farther by sandy-shale and shale turbidites of the Dzhemkukan Formation. The latter seems to be formed in a deep sea at the periphery of the major submarine fans, since locally (e.g., in the midreaches of the Bol'shoi Patom River) they are truncated by erosion channels filled with shallow-water clastic rocks, oncolite limestones, and turbidites.

Signs of iceberg rafting in Dzhemkukan shale facies are rare (Bol'shoi Barakun and Kulikach rivers). It implies that the rafting was limited to the margins of the Patom basin. It could have been possible if the glaciers had a "warm lining" (about 0°C) and the moraine material melt out almost entirely from the ice at the line of the glaciers' floating, or if the Patom basin featured high tides that broke off icebergs in shallow zones and did not allow them to drift. It is possible that both conditions were present in the Patom basin at the time of Middle Siberian glaciation. This hypothesis is corroborated by the submarine fans widespread in the Middle Siberian Glaciohorizon.

For all their fragmentary patterns, the available paleogeographic data strongly suggest that at the time of the Middle Siberian glaciation, the source areas and hence, the glaciation centers were situated on the Siberian Platform east and northwest of the Patom Depression. The extensive areas involved in glaciations, and the fact that glaciers formed a broad front descending all the way to the sea level, are suggestive of the cover-like nature and the regional scale of the Middle Siberian glaciation. Should the assumption of the Dzhemkukan age and the glacial genesis of the diamicts in the Southern Pribaikal'e zone be proved, the regional scope of the glaciation will be even more evident.

MIDDLE SIBERIAN GLACIOHORIZON AND LATE PRECAMBRIAN GLACIATIONS

Due to the uncertain age of the Mid-Siberian Glaciohorizon, it is impossible to specify its place in the row of Late Precambrian glaciations, which has been termed in the glacial geology literature, the African Glacial Era (Chumakov, 1978), or the Late Proterozoic

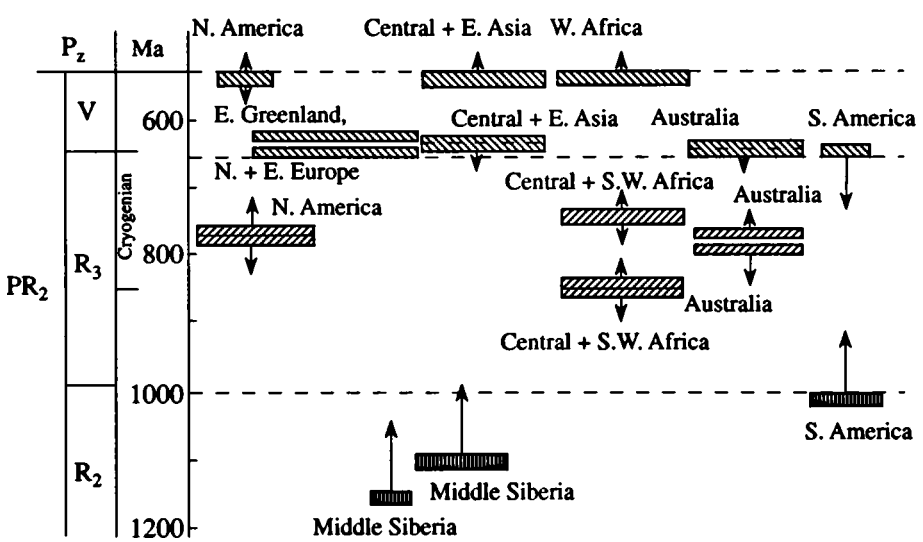


Fig. 6. *Stratigraphic position of the Mid-Siberian Glaciohorizon relative to other Late Precambrian glacial horizons. The arrows signify the uncertainty of age limits.*

Glacial Era (Hambrey and Harland, 1985). Despite the uncertainty, some speculation can be ventured. It has long been known that there were at least two glaciations in both the Vendian and Late Riphean (Chumakov, 1978; Hambrey and Harland, 1985). The Late Vendian glaciation, almost at the Cambrian boundary, seemed comparatively small (Central Asia and Kazakhstan: the Baikonur Glaciohorizon; Northern China: the Luoquan Glaciohorizon and its analogues). Recent finds of Lower Cambrian small-shelly fossils in the cap dolomites and limestones of the vast West African Glaciohorizon (Culver *et al.*, 1988) show that, first, this glaciation encompassed the lowermost Cambrian, and, second, it must have been more extensive than previously believed. Note that Lower Cambrian tillites were assumed to be present also in the Upper Tindir Group of Alaska (Allison *et al.*, 1981), and in Hoggar in Northwestern Africa (Caby and Fabre, 1981a).

Other data that refine our knowledge of the structure of the African Glacial Era have been found in South America. A Late Precambrian glacial horizon widespread over the San Francisco craton, with an age around 1010 - 1080 Ma has been identified there (D'Agrella-Filho *et al.*, 1990). The dating involves certain geological assumptions. If proved, it will imply that in South America major Riphean glaciations occurred at the end of the Middle Riphean, and that the African Glacial Era started much earlier than formerly believed. Its upper and presumably lower boundaries stretch much farther than the limits of the Cryogenian suggested by the International Subcommission on Precambrian Stratigraphy as one of the general divisions of the Neoproterozoic (Fig. 6).

In this light, the Middle Siberian Glaciohorizon is of particular interest as it may also have a Middle Riphean age. It is not excluded that the Middle Siberian Glaciohorizon, which may not even be from the latest Middle Riphean, will be evidence of the earliest Riphean glaciation to be found on Earth, or one of the earliest. Some reservation is needed here, since the nature of the diamictic sequences that occur in the lower Ballaganakh Group of the Patom Supergroup and in the Medvezhevka Formation of the Teptorgo Group (see Table 1) is not clear. As a result, specifying the age of the Middle Siberian Glaciohorizon and the origin of Ballaganakh and Teptorgo diamicts is a task important not only for the regional geology but for the Precambrian stratigraphy and historical geology in general. Even if the age of the Middle Siberian Glaciohorizon is proved to be Late Riphean, the implications will nevertheless be of importance for the Precambrian stratigraphy by presenting a first record of Late Riphean glaciations in Eurasia.

REFERENCES

Aitken, J.D., The Ice Brook Formation and Post-Rapitan, Late Proterozoic Glaciation, Mackenzie Mountains Northwest Territories, *Bull. Geol. Surv.*, Canada, 1991, no. 404, p. 43.

Allison, C.W., Young, G.M., Yeo, G.M., and Delaney, G.D., Glacigenic Rocks of the Upper Tindir Group, East-Central Alaska, *Earth's pre-Pleistocene glacial record*, Cambridge University Press, 1981, pp. 720 - 723.

Artyushkov, Ye.V., On the Establishment of Isostatic Equilibrium of Earth Crust, *Izv. Akad. Nauk SSSR, Ser. Fizika Zemli*, 1967, no. 1, pp. 3 - 16.

Belogur, V.D. and Tikhonov, V.L., Correlation of Baikal and Patom Supergroups at Their Conjunction Zone, *Voprosy geologii Pribaikaliya i Zabaikaliya*, 1970, no. 7, part 1, Chita, Trans-Baikal Branch of the USSR Geographic Society, 1970, pp. 3 - 8.

Bobrov, A.K., *Straigrafiya i Paleogeografiya Otlozhenii Verkhnego Dokembriya Yuzhnoi Yakutii* (Straigraphy and Paleogeography of Upper Precambrian Strata in Southern Yakut), Yakutsk, 1979.

Bozhko, N.A., On the Presence of Tilloids in the Precambrian Sequence in the Central Part of the Western Pribaikal'e, *Vestnik MGU, Geologiya*, 1976, no. 4, pp. 106 - 108.

Caby, R. and Fabre, J., Tillites in the Latest Precambrian strata of the Touareg Shield (Central Sahara), *Earth's Pre-Pleistocene Glacial Record*, Cambridge University Press, 1981a, pp. 146 - 149.

Caby, R. and Fabre, J., Late Proterozoic to Early Paleozoic Diamictites, Tillites and Associated Glacigenic Sediments in the Serie Pourpree of Western Hoggar, Algeria, *Earth's Pre-Pleistocene Glacial Record*, Cambridge University Press, 1981b, pp. 140 - 145.

Chumakov, N.M., Straigraphy and Tectonics of the Southwestern Part of the Vilui Depression, *Tektonika SSSR*, AN SSSR, Moscow, 1959, vol. 4, pp. 345 - 460.

Chumakov, N.M., Vendian Glaciation in Europe and the North Atlantic (Upper Precambrian), *Dokl. Akad. Nauk SSSR*, 1971, vol. 198, no. 2, pp. 419 - 422.

Chumakov, N.M., *Dokembriiskie Tilly i Tilloidy* (Precambrian Tills and Tilloids), Nauka, Moscow, 1978.

Chumakov, N.M., *Izucheniye Drevnikh Lednikovykh Otlozhenii, Prakticheskoe Rukovodstvo* (Studies of Ancient Glacial Strata, Practical Guide), *GIN Akad. Nauk SSSR*, Moscow, 1990.

Chumakov, N.M., Middle Siberian Glacial Horizons: Traces of the Earliest Late Precambrian Glaciation?, Earth's Glacial Record, *Abstracts of Project 260 TGCO*, 1991, p. 9.

Chumakov, N.M. and Krasil'nikov, S.S., Lithological Properties of Riphean Tilloids of the Ura Rise (Lena River), *Litologiya i Poleznye Iskopaemye*, 1991, no. 3, pp. 57 - 78.

Chumakov, N.M. and Semikhatov, M.A., Riphean and Vendian of the USSR, *Precambrian Res.*, 1981, vol. 15, pp. 229 - 253.

Cowie, J., Zieglor, W., and Remane, J., Stratigraphic Commission Accelerates Progress. 1984 to 1989, *Episodes*, 1989, vol. 12, no. 2, pp. 79 - 83.

Culver, S.J., Pojeta, J., and Repetski, J.S., First Record on Early Cambrian Shelly Microfossils from West Africa, *Geology*, 1988, vol. 16, no. 7, pp. 596 - 599.

D'Agrella-Filho, M.S., Pacca, I.G., Teixeira, W., Onstott, T.G., and Renne, P.P., Paleomagnetic Evidence for Evolution of Meso- to Neo-Proterozoic Glaciogenic Rocks in Central-eastern Brazil, *Palaeogeog., Palaeoclim., Palaeoec.*, 1990, vol. 80, no. 3/4, pp. 255 - 265.

Dol'nik, T.A. and Vorontsova, G.A., *Biostratigrafiya Verkhnego Dokembriya i Nizhnikh Gorizontov Kembriya Severo-Baikal'skogo i Patomskogo Nagorii* (Biostratigraphy of the

Upper Cambrian and Lower Cambrian Horizons of the Northern Baikal and Patom Uplands), Eastern Siberia, Irkutsk, 1974.

Dol'nik, T.A., Mitrofanov, G.L., Taskin, A.P., Stetsyuk, M.I., and Butin, K.S., A Scheme of the Stratigraphy of Upper Precambrian Strata in the Fold Belt of Southeastern Siberia, *Straigraficheskoye i Paleontologicheskoe Obosnovanie Legend Geologicheskogo Kartirovaniya v Masshtabe 1:50000 Vostochnoi Sibiri*, SNIIGGiMS, Irkutsk, 1987, pp. 93 - 99.

Dreimanis, A., The Problem of Waterlain Tills, *Moraines and Varves*, Schuchter, Ch., Eds., 1979, pp. 167 - 179.

Golovenok, V.K., Salop, L.I., and Chumakov, N.M., Northern Part of the Baikal Mountains Area, *Verkhnii Dokembrii*, AN SSSR, Moscow, 1963, pp. 436 - 458.

Gravenor, C.P., Von Brunn, V., and Dreimanis, A., Nature and Classification of Waterlain Glaciogenic Sediments, Exemplified by Pleistocene, Late Paleozoic and Late Precambrian Deposits, *Earth. Sci. Rev.*, 1984, vol. 20, pp. 105 - 166.

Hambrey, M.J. and Harland, W.B., The Late Proterozoic Glacial Era, *Paleogeog., Palaeocl., Palaeoec.*, 1985, vol. 51, no. 1 - 4, pp. 255 - 272.

Hambrey, M.J., Harland, W.B., Chumakov, N.M., Fancelt, K.E., and Herod, K.N. *Earth's pre-Pleistocene Glacial Record*, Wang Zejin, Eds., Cambridge Univ. Press, Cambridge, 1981.

Harland, W.B., Herod, K.N., and Krinsley, D.H., The Definition and Identification of Tills and Tillites, *Earth Science Reviews*, 1966, vol. 2, pp. 225 - 256.

Keller, B.M., Semikhatov, M.A., and Chumakov, N.M., The Upper Proterozoic of Siberian Platform and Its Framing, *Straigrafiya Dokembriya i Kembriya Srednei Sibiri*, Krasnoyarsk, 1967, pp. 247 - 291.

Khomentovskii, V.V., Baikalium as the Principal Stage in the History of Siberia's Geological History, *Geologiya i Geokhronologiya Dokembriya Sibirskoi Platformy i Ee Obramleniya*, Nauka, Leningrad, 1990, pp. 222 - 237.

Khomentovskii, V.V., Shenfil', V.Yu., Yakshin, M.S., and Butakov, E.P., *Opornye Razrezy Otlozhenii Verkhnego Dokembriya i Nizhnego Kembriya Sibirskoi Platformy* (Key Sections of Upper Precambrian and Lower Cambrian Strata on the Siberian Platform), Nauka, Moscow, 1972.

Krasil'nikov, S.S., New data on the Structure of the Riphean in the Northwestern Slope of the Chuya Anticlinorium, *Regional'naya Geologiya: Tektonika, Litologiya i Stratigrafiya*, GIN Akad. Nauk SSSR, Moscow, 1990, pp. 68 - 74.

Link, P.K. and Gostin, V.A. Facies and Palaeogeography of Sturtian Glacial Strata (Late Precambrian), South Australia, *Amer. J. Science*, 1981, vol. 281, pp. 353 - 374.

Mats, V.D. and Bukharov, A.A., To the Stratigraphy of the Upper Precambrian of the Western Margin of the Baikal Mountains, *Straigrafiya Dokembriya i Kembriya Srednei Sibiri*, Krasnoyarsk, 1967, pp. 339 - 357.

Pavlovskii, Ye.V. and Belichenko, V.G., Western and Eastern Peri-Baikal, *Verkhnii Dokembrii*, AN SSSR, Moscow, 1963, pp. 401 - 434.

Pettijohn, F.J., *Sedimentary Rocks*, Harper and Bros, New York, 1957.

Preiss, W.V., The Adellide Geosyncline. Late Proterozoic Stratigraphy, Sedimentation, Paleontology and Tectonics, *Bull. Geol. Surv. S. Aust.*, 1987, vol. 53, p. 458.

Salop, L.I., *Geologiya Baikal'skoi Gornoi Oblasti* (Geology of Baikal Mountains), Nedra, Moscow, 1964, vol. 1.

Sharov, V.N., Fefelov, N.N., Zarudneva, N.V., Rusakova, V.A., and Brant, S.B., Pb-Pb Dating of Crystalline Schists of the Mam Synclinorium (Baikal-Patom Upland), *Dokl. Akad. Nauk SSSR*, vol. 319, no. 1, pp. 209 - 212.

Stanevich, A.M., Dol'nik, T.A. and Faizulina, Z.Kh., Major Boundaries of Changes in Various Fossil Groups of the Precambrian of Southern East Siberia, and Their Age Implications, Stratigraphy of the Upper Proterozoic of the USSR (Riphean and Vendian), *II Vseosyuznoye Soveshchaniye "Obshchie Voprosy Raschleneniya Dokembriya SSSR"* (2nd All-Union Conf. "General issues of USSR Precambrian Stratification"), Abstracts of Papers, Bashkirgeologya, Ufa, 1990, pp. 79 - 81.

Zubtsov, E.I., Precambrian Tillites of the Tien Shan and Their Stratigraphic Implication, *Bull. MOIP. Otd. Geol.*, 1972, no. 1, pp. 42 - 56.

Stratigraphy and Geological Correlation, Vol. 1, No. 1, 1993.Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 1, 1993, pp. 35 - 40.
Original Russian Edition Copyright © 1993 by Stratigrafiya. Geologicheskaya Korrelyatsiya, Bibikova, Gracheva, Makarov, Nozhkin.
English Translation Copyright © 1993 by Interperiodica.

The Age Boundaries of the Early Precambrian Geological Evolution of the Enisei Ridge

E. V. Bibikova*, T. V. Gracheva*, V. A. Makarov*, A. D. Nozhkin**

**Vernadskii Institute of Geochemistry and Analytical Chemistry, Ul. Kosigina 19, Moscow, 117975, Russia*

***Institute of Geology and Geophysics, Siberian Branch of the Russian Academy of Sciences, Universitetskii Pr. 3, Novosibirsk, 630090 Russia*

Manuscript received May 15, 1991

Abstract – U-Pb isotopic dating was made for the accessory zircons from the granulites of the Kan Group and the plagiogneiss of the Enisei Group of the Enisei Ridge. The accumulation of the Enisei Group (1880 ± 20 Ma.), the metamorphism of granulite facies within the Kan Group (1900 ± 10 Ma.), and the rock granitization of both groups (1780 ± 20 Ma.) took place in the Early Proterozoic. An age of 2730 ± 180 Ma., corresponding to the Late Archean in the stratigraphic scale, was determined for the protolith of the Kan rocks.

The Enisei Ridge is a Precambrian folded region that forms part of the marginal structures of the southwestern Siberian Platform. It stretches along the Enisei River as a stripe 100 to 200 km wide and over 700 km long. It is composed of metamorphic, magmatic, and sedimentary rocks of the Archean to the Phanerozoic. The Kan and Enisei groups are the oldest geological formations of the Enisei Ridge and are distributed in the southern part of the ridge within the Angaro-Kan crystalline basement prominence.

The structural features and high-grade metamorphism of these rocks have been used by some authors to refer these formations to the oldest early Archean complexes on Earth. This was supported by some isotopic datings, which, however, were not always obtained from the correct isochron constructions: an Rb-Sr age of 2.55 ± 0.1 Ga., with isochron for two whole-rock points (Gerling, Artemov, 1964) and a U-Pb zircon age of 4.1 ± 0.3 Ga., in which common lead content exceeded 90% (Volobuev *et al.*, 1980).

The rocks of the Kan Group that metamorphosed in granulite facies conditions are present at the base of the Precambrian sequence and compose most of the prominence area. They are represented by pyroxene-bearing plagiogneiss of dacite-rhyodacite and occasionally, andesite composition, as well as by aluminiferous gneiss and two-pyroxene schists of basaltic and basaltic-andesite composition, which are distributed in the lower part of the sequence. Auto- and allochthonous charnockitoids (bogunites and kuzeevites according to Yu.A. Kuznetsov, 1941) are quite common. The charnockite-granulite complex was concluded by the formation of leucocratic orthoclase granite and pegmatoids. A major autochthonous granitoid pluton (Tarak) is located in the rocks of the northeastern part of the complex. Together with gneiss, it forms a single structure with gradual transitions from granites to granulites through the zone of migmatized gneiss.

The original nature of the metamorphic rocks of the Kan Group is most probably volcano-sedimentary: two-pyroxene schists and plagiogneiss are orthorocks, being the products of contrasting volcanism, while aluminiferous gneiss belongs to pararocks.

A distinctive geochemical feature of the Kan Group rock is a higher concentration in gneiss of such low-mobility elements as TR, Th, and Zr. Moreover, lowered uranium concentrations are certainly due to the removal of uranium in the course of the metamorphism of granulite facies, which resulted in high Th/U ratios (15 - 50).

The heterogeneity of the metamorphism of the Kan rocks appears to be its most typical feature. It is due to the block structure of the crystalline prominence and the variable depth of the erosion of the sequence, exceeding 6.5 to 7 km in total thickness (which corresponds to the pressure difference of 2.5 kbar). Reaction structures and chemical features of minerals in the Kan rock complex generally indicate retrograde metamorphism. Nevertheless, judging from the paragenesis development of Hyp + Sill, Qz + Cord + Garn + Sill, and eclogite-like rocks, the pressure at metamorphism reached from 9 to 10 kbar and the temperature was not less than 850 - 950°C.

On the basis of mineral zonation and calculations of fluid composition, two stages of retrograde granulite- and amphibolite-facies metamorphism have been identified: the earlier stage (800 - 650°C) was characterized by the hydrous composition of the metamorphic fluid and gentle *T–P* cooling trends, and the late stage (650 - 550°C) was characterized by almost purely carbonic acid fluid composition. The latter stage corresponds to the growth of granite-gneissic domes against the background of an intensive inflow of CO_2 fluid (Gerya, 1989; Perchuk *et al.*, 1989). Later retrograde metamorphism in epidote-amphibolite facies (550 - 450°C, 3 - 2 kbar) has also been established.

A U-Pb isotopic study was made for the accessory zircons from the three granulite samples of the Kan Group rock.

The A-204-85 Sample is composed of fine-grained hypersthene plagiogneiss, collected near the mouth of the Shilka River (Shilka block). It is a massive, slightly banded rock with a granoblastic texture. It is predominantly composed of plagioclase and quartz, but containing hypersthene (10 - 15%), biotite (up to 4%), magnetite, an ilmenite admixture, accessory zircon, and apatite as well. The chemical composition of gneiss (Table 1) corresponds to that of plagiorhyodacite.

The accessory zircon in the sample is mainly represented by the generation of small, isometrical, multifaceted grains, which are sometimes partly melted. Grains are transparent, colorless, and lustrous. This agrees well with the morphological type of zircon formed by solid-phase crystallization under conditions of the metamorphism of granulite facies (Fig. 1, II). About one percent of the dark brown, prismatically-dipyramidal opaque crystals is present.

The A-304-85 Sample consists of fine-grained, hypersthene plagiogneiss, collected in the middle reaches of the Nemkina River (Kuzeev block). The main rock mass of the sample is a lenticular aggregate with fine (up to 1.5 mm) hypersthene grains forming elongated glomeroblastic accumulations in the quartz-feldspathic groundmass. Other minerals include garnet, biotite, accessory apatite, zircon, magnetite, and ilmenite. The patches composed of an hypersthene-quartz-orthoclase aggregate, grading into fine-grained mass, are sometimes observed in the gneiss. Such aggregates were formed by recrystallization under conditions similar to the process of the initial stage of charnockite formation. The chemical composition of the samples is given in Table 1.

The accessory zircon in the sample is heterogeneous. Three varieties have been identified: prismatic grains, which are brown and opaque with a strongly corroded surface; prismatic, translucent grains with smoothed contours, which are highly fractured; and isometrical, slightly colored, transparent grains. Only the third variety corresponds to the "granulitic" type (Fig. 1). The first type, most probably, corresponds to the relicts of the originally magmatic zircon.

The 156 Sample, which is composed of enderbite (bogunite), was collected on the left side of the Enisei River valley, 250 m upstream from the Savina River from a sheet-like body, which forms part of a series of alternating garnet and hypersthene gneiss and two-pyroxene basic schists. The rock is medium- to coarse-grained and lenticular-banded: the more melanocratic bands are enriched with garnet or hypersthene (sometimes with a biotite admixture), while the leucocratic bands are enriched with plagioclase having a minor amount of quartz. The main minerals are plagioclase (sometimes with antiperthite K-feldspar ingrowths), poorly pleochroic hypersthene, and garnet containing 6.27% MgO, 4.94% CaO and 28.24% FeO. Brownish biotite (TiO_2 – 6.2%, FeO – 14.05%, MgO – 14.42%) develops after hypersthene and garnet; it is commonly associated with fine-grained quartz in which zircon is commonly present. An ore mineral and apatite occur together with zircon as an admixture. The chemical rock composition corresponds to diorite.

Accessory zircon is represented by prismatic dipyramidal grains with smoothed melted edges. The zircon is slightly colored, transparent and translucent, corresponding to the zircon of the morphological type,

Table 1. Chemical composition of studied rocks (wt. %)

Components	A-204-85	A-304-85	156	A-277-85	112
SiO_2	69.03	69.14	54.77	63.59	62.35
TiO_2	0.73	0.85	0.41	0.81	1.11
Al_2O_3	13.30	12.31	16.11	14.50	14.02
Fe_2O_3	2.68	0.90	2.69	8.32	1.72
FeO	3.62	5.90	9.75	–	7.37
MnO	0.07	0.08	0.16	0.07	0.16
MgO	2.45	2.96	4.31	2.96	3.27
CaO	3.15	3.48	6.40	2.92	4.04
Na_2O	2.80	2.04	2.53	3.33	2.07
K_2O	1.24	1.56	1.00	2.48	2.39
P_2O_5	0.05	0.13	0.29	0.10	0.09
Loss on ignition	0.50	0.56	1.62	1.00	1.41
Total	99.72	100.00	100.04	100.14	100.00
U(g/t)	0.7	1.1	0.2	1.0	0.2
Th(g/t)	11.8	7.5	0.3	27.2	4.0
Th/U	17.0	6.8		27.2	20.0

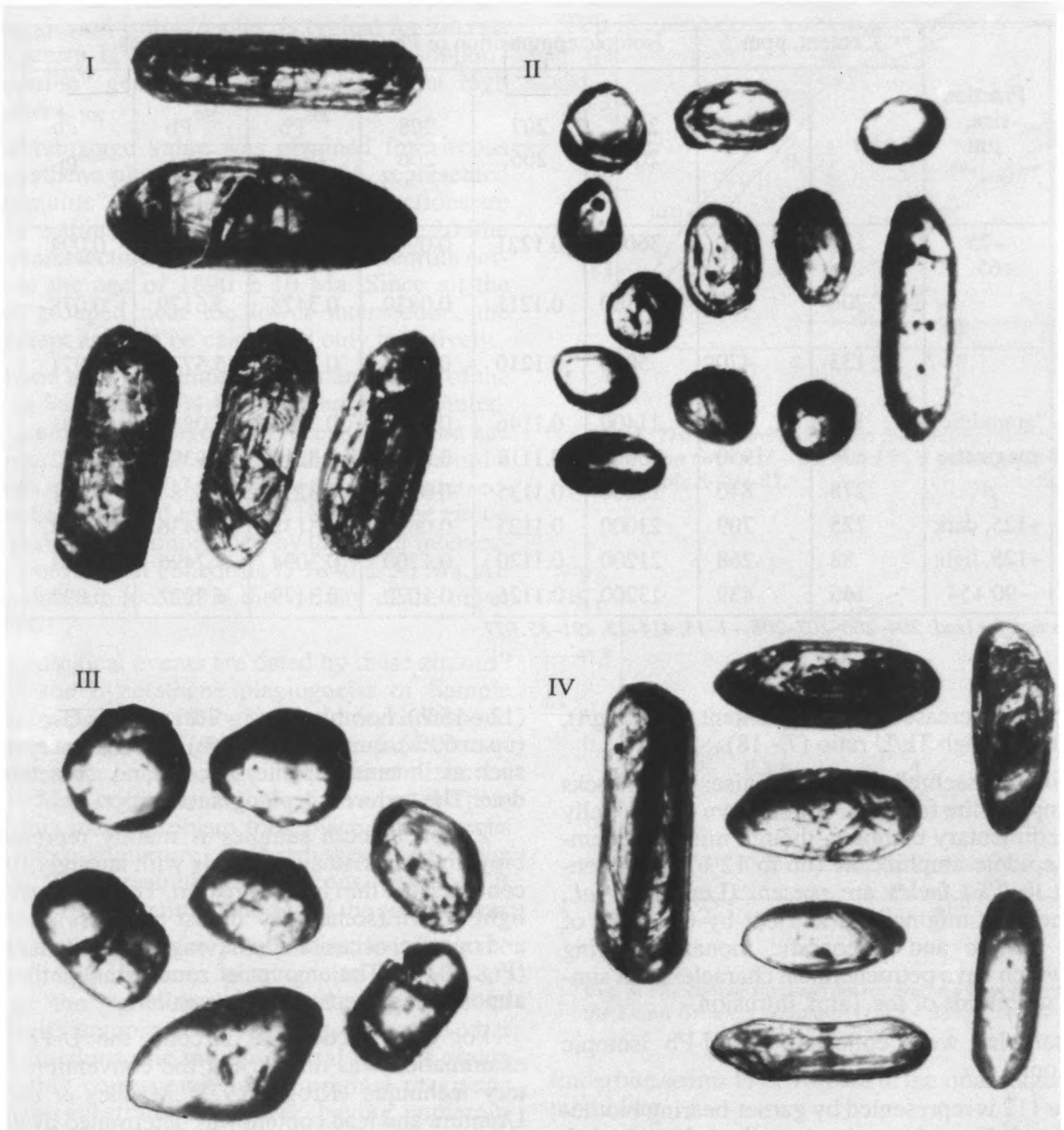


Fig. 1. *The morphology of zircons from Kan and Enisei groups of the Enisei Ridge.*
I – Sample 156, fraction +90, II – Sample A-204-85, fraction +75, III – Sample A-304-85, fraction –65, IV – Sample A-277-85, fraction +100.

which crystallized from the melt under conditions of the metamorphism of granulite facies (Fig. 1, I).

The Enisei Group rocks (volcanogenic, carbonate, and terrigeneous) metamorphosed in the amphibolite facies stretch along the Enisei River as a narrow stripe (1 - 1.5 km wide), contacting with the granulites of the Kan Group along the mylonite zone. In addition, they are distributed in the northern part of the crystalline basement prominence, overlying the rocks of the granulite facies. The migmatite-gneiss complex of the Enisei Group is composed of garnet-biotitic (less frequently, garnet-biotite-sillimanitic paragneisses), potassic and potassium-sodic microgneiss (acid metavolcanic rocks), biotite-amphibole plagiogneisses (meta-andesites), and amphibolites (metabasites). Metamorphosed bimodal volcanic rocks are present in the lower part of the sequence together with subordinated pararocks. Marble, calciphyre, and amphibolite (metabasite) bodies occur in the upper part of the sequence alongside with gneiss. The presence of acidic volcanic rocks with an increased radioactive element content (U = 3 - 4.5 g/t, Th = 20 - 45 g/t, and K = 3.5 - 4.0%) testifies to the manifestation of acidic crustal magmatism in that epoch and to the greater thickness and differentiation of the underlying protocontinental crust. Small-size elastic metarocks are also characterized by a high thorium content

Table 2. Isotopic age of zircons from the rocks of the Kan Group

Sample No.	Fraction size, μm	Content, ppm		Isotopic composition of Pb			Isotopic ratios			$\frac{^{207}Pb}{^{206}Pb}$ age, Ma
		Pb	U	$\frac{206}{204}$	$\frac{207}{206}$	$\frac{208}{206}$	$\frac{^{206}Pb}{^{238}U}$	$\frac{^{207}Pb}{^{235}U}$	$\frac{^{208}Pb}{^{206}Pb}$	
A-204-85	–75 +65	177	490	36000	0.1221	0.0948	0.3448	5.7865	0.094	1981
	– 65	206	578	6100	0.1211	0.0839	0.3428	5.6179	0.078	1939
	+7 5	133	470	5000	0.1210	0.0740	0.3414	5.5723	0.071	1931
A-304-85	"granulitic"	168	498	11400	0.1146	0.0938	0.3253	5.0882	0.091	1855
	magmatic	636	1950	25600	0.1119	0.0632	0.3218	4.9397	0.062	1821
	»	278	840	15000	0.1135	0.0735	0.3237	5.0284	0.071	1842
156	+125, dark	225	709	23000	0.1125	0.0651	0.3138	4.8436	0.064	1831
	+125, light	88	268	21200	0.1120	0.1200	0.3094	4.7496	0.118	1821
	–90 +54	146	439	13200	0.1126	0.1020	0.3179	4.8922	0.099	1826

Note: Correction for lead: 204–206–207–208 – 1–15, 424–15, 291–35, 077.

(15 - 30 g/t), a decreased uranium content (2 - 3.2 g/t), and a relatively high Th/U ratio (7 - 18).

The mineral assemblages of the Enisei Group rocks suggest amphibolite facies metamorphism of originally volcano-sedimentary complex, though mineral assemblages of epidote-amphibolite (up to 12%) and greenschist (up to 7%) facies are present (Lepezin *et al.*, 1981). Gneiss is migmatized and cut by the veins of gneissose biotitic and leucocratic monazite-bearing granites, which have petrochemical characteristics similar to the granitoids of the Tarak intrusion.

Two samples were collected for U-Pb isotopic examination:

Sample 112 is represented by garnet-bearing biotite-amphibolic plagiogneiss and was collected on the left side of the Enisei River valley 0.6 km downstream from the Osinovyi Creek from a member of gneiss of dacitic composition intercalated with amphibolite. It is fine-grained, thin-banded rock and contains garnet segregations (up to 1.5 mm). The mineral assemblage is: biotite (15 - 17%), hornblende (15 - 20%), garnet (10%), plagioclase (andesine) (40 - 50%), quartz (10 - 15%), and zircon less than (<1%). By its chemical composition (Table 1), the rock corresponds to andesite, though the original structure of the volcanic rock was not preserved.

Sample A-277-85 is amphibole-biotitic gneiss collected on the right side of the Enisei River valley 0,6 km downstream from the village of Predivinsk from a member of alternating migmatized biotitic or biotite-amphibolic gneiss with layers of amphibolite and amphibole plagiogneiss. It is a fine-grained, relatively homogeneous or slightly banded rock, consisting of biotite (12 - 15%), hornblende (5 - 7%), plagioclase (andesine) (up to 60%), quartz (up to 15%), and accessory minerals, such as ilmenite, apatite, zircon, and, sometimes, epidote. The texture is lepidogranoblastic.

Zircon in both samples is mainly represented by bipyramidal prismatic crystals with strongly smoothed contours and thin inner zonation. The grains are dark to light brown (sometimes almost colorless), translucent and opaque, occasionally having dark dotted inclusions (Fig. 1, IV). The amount of rounded, slightly colored, almost transparent grains is smaller.

For size fractions of zircons, the U-Pb isotopic examination was done using the conventional laboratory technique (Krogh, 1973; Manhes *et al.*, 1978). Uranium and lead content was determined by the isotopic dilution method using a mixed isotopic tracer ($^{208}Pb + ^{235}U$).

The measurement accuracy of U and Pb concentrations is ± 1%. The lead isotopic composition was measured by a TSN-206A Cameca mass-spectrometer using silica gel as an ion emitter.

The accuracy of the isotopic ratio measurements is ± 0.1%. Isotopic ages were calculated using standard values of decay constants (Steiger, Yager, 1976). The calculation of the isochron relations was done using the least squares technique; the sum of the squared deviations of the data point from the straight line was minimized. The errors are given at a confidence level of 95%.

The results obtained are shown in Table 2 and on the concordia diagram (Fig. 2).

A characteristic geochemical feature of zircons from the Kan rocks is their extremely low content of

common lead. This feature, as well as the high concordance of evaluated isotopic ages, is typical for zircons whose structure is highly crystalline. This supports their "granulite" genesis and crystallization at high PT-parameters.

A maximum age value was obtained for zircons from hypersthene plagiogneiss A-204-85, represented by one "granulite" generation. Two zircon fractions are concordant within 10%, their age being 1920 ± 20 Ma. The lower intersection of discordia with concordia corresponds to the age of 1890 ± 10 Ma. Since all the points are grouped near the lower intersection, the upper intersept age can be calculated only tentatively.

The zircon ages are almost concordant in one of the fractions in Sample A-304-85 (dark brown, fractured, corroded grains). Their age is 1810 ± 10 Ma. The age by the lower intersection of discordia with concordia corresponds to 1780 ±10 Ma, and by the upper intersection, it can be estimated as 2730 ± 180 Ma. The zircon age from enderbite (Sample 156) by the upper intersection of discordia with concordia is 1840 ± 50 Ma. All the data points are located in the vicinity of the upper intersection.

What geological events are dated by these zircons? Zircons in the hypersthene plagiogneiss of Sample A-204-85, by their morphological features, correspond to the zircons originated at the granulite-facies metamorphism stage. The age of these zircons (1900 ± 20 Ma.) corresponds to the time when the processes of this metamorphism took place. The zircons contain a micro-admixture of old radiogenic lead of the substratum. This assemblage of zircons, however, is not adequate for determining the age of the substratum.

Only one of the accessory zircon fractions from the sample of hypersthene plagiogneiss (A-304-85) belongs to the "granulite" type. By the ratio of $^{207}Pb/^{206}Pb$, its isotopic age is 1855 Ma. In the two other examined fractions, the morphological type of zircon most probably corresponds to the original magmatic zircon of the substratum, however, having numerous indications of recrystallization and solution, that probably reflect the granitizing effect of the Tarak granite intrusion, in the vicinity of which the above sample was collected. The U-Pb monazite and zircon age of these granites is 1780 ± 50 Ma (Geokhronologiya, . . . 1973). The same age of 1780 ± 20 Ma was determined for zircons from Sample A-304-85 by the lower intersection of discordia with concordia. The age of the relict substratum component was evaluated by the upper concordia intersection with a large experimental error of 2730 ±180 Ma.

The accessory zircons in enderbite (Sample 156) (by their morphological features) correspond to zircons formed from the melts under conditions of granulite facies metamorphism. The age of 1840 ± 50 Ma, obtained using these zircons, is somewhat lower then, but with due account for the experimental error it does not contradict the age of 1900 ± 20 Ma that we have calculated for the granulite facies metamorphism.

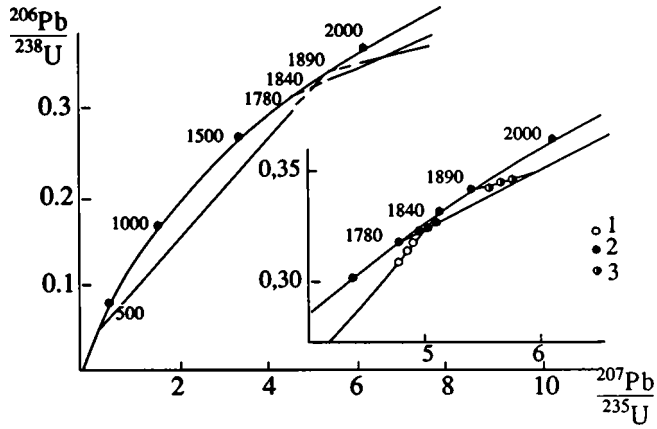


Fig. 2. *The concordia diagram for zircons from the rocks of the Kan Group. 1 – Sample 156, 2 – Sample A-304-85, 3 – Sample A-204-85.*

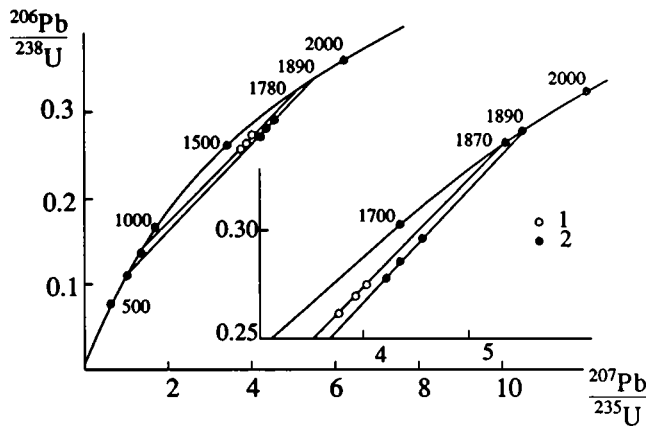


Fig. 3. *The concordia diagram for zircons from the rocks of the Enisei Group. 1 – Sample 112, 2 – Sample A-277-85.*

Enderbite seems to be formed at the final stage of metamorphism.

The geochemical features of zircons from the rocks of the Enisei Group (particularly, Sample 112) indicate their magmatic genesis in rocks of volcanic origin. These features include a relatively low uranium content (the ratio $(^{208}Pb/^{206}Pb)_{rad}$ value *is ca.* 0.1) (Table 3), and the higher common lead content in comparison to that of the "granulite" zircons. By the upper intersection of discordia with concordia, the age for Sample 112 is 1872 ± 24 Ma and for Sample A-277-85 – 1890 ± 20 Ma. The ages determined by the lower intersection are within the interval of 700 - 800 Ma. This may reflect a real event in the Riphean history that resulted in the partial removal of radiogenic Pb from the zircons (Fig. 3).

Because the rocks of the Enisei Group are considered to be metavolcanics and the originally magmatic genesis of zircons is evident, the age of their formation (1880 Ma) can be regarded as the time when the volcanogenic Enisei Group accumulated.

Table 3. Isotopic age of the rocks of the Enisei Group

Sample No.	Fraction size, μm	Content, ppm		Isotopic composition of Pb			Isotopic ratio			$\frac{^{207}Pb}{^{206}Pb}$ age, Ma
		Pb	U	$\frac{206}{204}$	$\frac{207}{206}$	$\frac{208}{206}$	$\frac{^{206}Pb}{^{238}U}$	$\frac{^{207}Pb}{^{235}U}$	$\frac{^{208}Pb}{^{206}Pb}$	
112	−100	48.0	166.5	3100	0.1113	0.1013	0.2748	4.0487	0.089	1746
	+125	62.5	226.1	3640	0.1086	0.1105	0.2622	3.7912	0.101	1712
	−125 +100	65.7	232.5	3425	0.1099	0.1073	0.2686	3.9226	0.097	1730
A-277-85	−125 +100	132.5	455.4	5960	0.1132	0.0653	0.2863	4.3795	0.059	1810
	+125	183.4	607.5	4100	0.1143	0.0666	0.2961	4.5786	0.0605	1832
	−100	148.5	523.1	4115	0.1134	0.0681	0.2781	4.2217	0.0600	1801

Note: Correction for lead: 204–206–207–208 – 1–15, 424–15, 291–35, 077.

Geochronological studies have made it possible to outline the following age boundaries of the geological evolution of the oldest rocks of the Enisei Ridge.

The age of the protolith of the Kan rocks is not less than 2730 ± 180 Ma. The available data is insufficient to establish the nature of this protolith.

Deep metamorphic transformations of the Kan rocks under conditions of granulite facies occurred 1900 ± 10 Ma ago, accompanied by charnockite formation 1840 ± 40 Ma ago, probably at the final stage of this metamorphic event.

The rocks of the Enisei Group seem to have formed in the Early Proterozoic, 1880 ± 20 Ma ago.

The most intense processes of granite formation within both groups took place 1780 ± 10 Ma ago, simultaneously with the emplacement of the Tarak granite intrusion.

REFERENCES

Geokhronologiya SSSR. Dokembrii (Geochronology of the Soviet Union. Precambrian), Nedra, Leningrad, 1973, vol. 1, pp. 189 - 202.

Geria, T.V., *Phase Correlation in Spinel-bearing Granulites*, Cand. Sci. (Geol.-Min.) Dissertation, Moscow University, Moscow, 1989.

Gerling, E.K., and Artemov, Yu.M., Absolute Geochronology of the Southern and Central Regions of the Enisei Ridge, *Geokhimiya*, 1964, no. 7, pp. 610 - 622.

Krogh, T.E., A Low-Contamination Method for Hydrothermal Decomposition of Zircon and Extraction of U and Pb for Isotopic Age Determination, *Geochim. et Cosmochim. Acta*, 1973, vol. 37, no. 3, pp. 485 - 500.

Kuznetsov, Yu.A., *The Precambrian Petrology of the South-Enisei Ridge. Materialy po Geologii Zapadnoi Sibiri* (Materials on the Geology of Western Siberia), West Siberian Gological Department, Tomsk, 1941, no. 15, (57).

Lepezin, G.G., Nozhkin, A.D., and Gerya, T.V., Thermodynamic Parameters of Metamorphism of the Kan Group, *Geologiya i Geofizika*, 1981, no. 6, pp. 11 - 19.

Manhes, G., Minster, J.F., and Allegre, C.J., Comparative U-Th-Pb and Rb-Sr Study of the Saint Severin Amphoterite, *Earth Planet. Sci. Lett.*, 1978, vol. 39, no. 1, pp. 14 - 27.

Perchuk, L., Gerja, T., and Nozhkin, A., Petrology and Retrograde P-T Path in Granulites of the Kanskaya Formation, Enisei Range, Eastern Siberia, *J. Metamorphic Geol.*, 1989, vol. 7, pp. 599 - 617.

Steiger, R., and Yager, E., Subcomission on Geochronology: Convention on the Decay Constants in Geo- and Cosmochronology, *Earth Planet. Sci. Lett.*, 1976, vol. 33, no. 3, pp. 359 - 362.

Volobuev, M.I., Zykov, S.I., Stupnikova, N.I., and Vorob'ev, I.V., Lead-Isotope Geochronology of the Precambrian Metamorphic Complexes of the South-Western Boundary of the Siberian Platform, *Geokhronologiya Vostochnoi Sibiri i Dal'nego Vostoka* (Geochronology of Eastern Siberia and Far East), Nauka, Moscow, 1980, pp. 14 - 30.

Stratigraphy and Geological Correlation, Vol. 1, No. 1, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 1, pp. 41 - 58.
Original Russian Edition Copyright © 1993 Stratigrafiya. Geologicheskaya Korrelyatsiya, Veis, Vorob'eva.
English Translation Copyright © 1993 by Interperiodica.

Evolution of the biosphere

Microbiotas of the Kerpyl' Group in the Riphean Hypostratotype of Siberia

A. F. Veis and N. G. Vorob'eva

Geological Institute, Russian Academy of Sciences, Pyzhevskii Per. 7, Moscow, 109017, Russia

Manuscript received June 13, 1992

Abstract – A succession of seventeen associations of organowalled and mineralized microfossils of different ages in deposits of the Kerpyl' Group (Uchur-Maya region, East Siberia) has been considered. The studied biotas are widely selected fossil microorganisms including representatives of all their main Precambrian morphological and ecological associations. Forty-four genera and 54 species of microfossils have been found. The distribution of organowalled microfossils in polyfacies deposits of the Kerpyl' group has been analyzed. By comparing Kerpyl' microfossils with biotas of the Riphean stratotype, we have confirmed our conclusion that Kerpyl' deposits refer to the Upper Riphean.

In recent years, significant progress has been made in the study of Precambrian microfossils. The stratigraphic range and area of observation have become wider. Many hundreds of microfossil genera and species distributed from the Upper Archean to the Proterozoic top have been described. Classification of microfossils has improved. Common features in their taxon regional successions are revealed. The limits within the successions are marked, and some demonstrate a wide stratigraphic value in their interregional consistency. Until now, there has been no general idea about the biostratigraphic potential of Precambrian microfossils and about their optimal use in subdivision and correlation.

One reason is that relatively poor, singular, or discretely distributed microbiotas are frequently involved in a biostratigraphic analysis. Some popular schemes of microbiotic systematic composition dependence on the conditions of its formation are speculative. Most importantly, there is no sufficient concrete data on real spatial-temporal distribution of microfossils even for regions thoroughly studied in the microphytologic respect.

This article is aimed at filling in the gap. It describes the distribution of organowalled and silicified microfossil biotas in a stratigraphically large fragment of the Riphean hypostratotype in Siberia, which is the Kerpyl' Group of the Uchur-Maya region (organowalled forms are mainly covered by our original data, silicified forms, by our literary data). Seventeen levels of age related microfossil occurrence were revealed within the Kerpyl' Group. They were observed in widely distributed layers differing in composition, facies appearance, and structural nature. What makes the Kerpyl' microbiotas especially interesting is their proximity to the most important historical, geological, and stratigraphic limit of the Upper Proterozoic, the boundary of the Middle and Upper Riphean.

Our method differed from the one described above (Veis, 1988; Veis, Semikhatov, 1989; Veis *et al.*, 1990) only by our selection of the largest microfossils to put under the binocular microscope without preliminary sediment centrifuging, and by our decision to centrifuge them in heavy liquid during maceration. This allowed us to add a number of unknown morphotypes to the sample of Kerpyl' microfossils, namely, especially large samples of acritarchs. We studied 340 samples and 120 of them revealed microfossils. Our original material in the following analysis was supplemented by literary data on all organowalled and silicified microfossils known from the Kerpyl' deposits (Veis, 1983; Golovenok, Belova, 1986; Pyatiletov, 1988; Timofeev *et al.*, 1976). When necessary, in accordance with the accepted classification of microfossils, revision of systematic composition was made for this data (*Mikrofossilii*, ... 1988).

Stratigraphic Position of Kerpyl' Microfossils

The Kerpyl' Group is the most widely distributed lithostratigraphic complex of the Uchur-Maya Riphean. Its outcrops can be observed in the Uchur, Maya, Yudoma, and Belaya basins and are confined to a number of large conjugated structural facies of the southeastern margin of the Siberian platform and

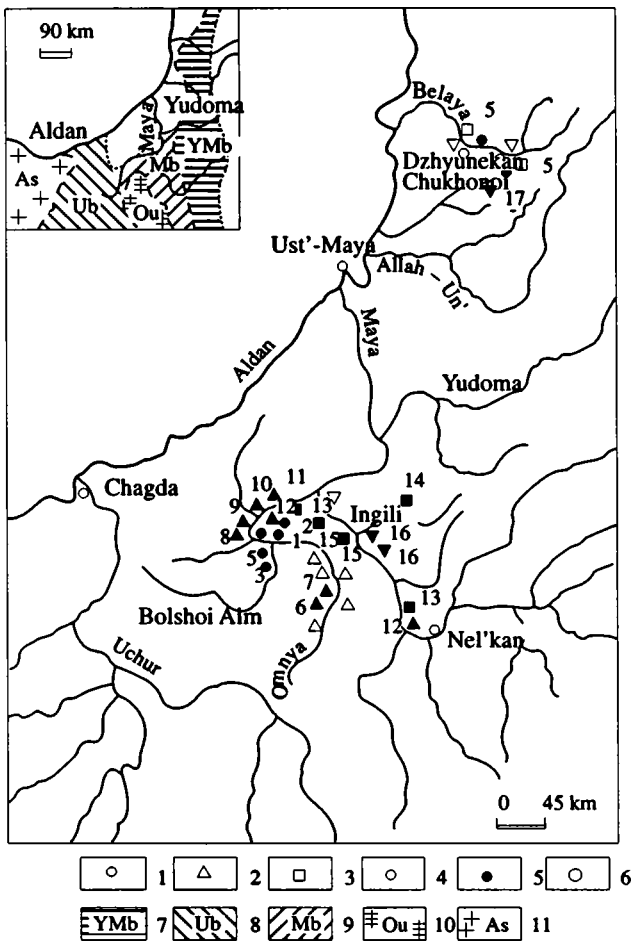


Fig. 1. *A schematized map of locations (levels of occurrence) of Kerpyl' Group microfossils in the Uchur-Maya region of East Siberia.*

1 - 4 – formations (subformations) of the Kerpyl' Group: 1, 2 – Totta Formation (1 – Konder Subformation, 2 – Omnya Subformation); 3 – Malgina Formation; 4 – Tsipanda Formation; 5, 6 points of sample collection: 5 – containing no microfossils; 6 – locations (levels of occurrence) of microfossils; 7 - 11 – the main structural elements of the Uchur-Maya region: 7 – Yudoma-Maya basin; 8 – Uchur basin; 9 – Maya basin; 10 – Omnya uplift; 11 – Aldan shield. Figures on the scheme are ordinal numbers of the levels of microfossil occurrence (see explanations in the text).

southwestern segment of the Verkhoyansk-Kolyma fold (Fig. 1). A scheme published in M.A. Semikhatov and S.N. Serebryakov's monograph (1983) has been used to subdivide the group.

Five large sedimentary rhythms identified as groups are known in a general succession of Riphean deposits. The Kerpyl' Group is the third from the bottom. On a larger part of the region it shows transgressive overlapping on different horizons of the Uchur and Aimchan underlying groups. In the arch of the largest positive structure of the region, the Omnya uplift, it passes onto the Archean basement of the Siberian craton. In western sections, Kerpyl' strata are separated from overlapping deposits of the Lakhanda group by a washout surface. Three formations are distinguishable within the group: (1) the Totta Formation, terrigenous mainly grey-colored (450 - 1500 m), (2) the Malgina Limestone variegated (50 - 440 m), and (3) the Tsipanda Dolomite grey-colored (200 - 420 m). In southwestern and northern Totta deposits, two subordinate units have been defined according to various criteria. These are Konder and Omnya, and Bik and Muskel' formations respectively. Stratigraphic positions of their boundaries do not coincide.

At present, the age of the Kerpyl' group is only roughly estimated. Some authors associate it either to the upper part of the Middle Riphean or with the lower part of the Upper Riphean. The boundary of the mentioned phytems is supposed to pass along the base or roof of the Kerpyl' Group, or in the upper part of the Lakhanda Group 9 (Komar, 1990; Semikhatov, Serebryakov, 1983; Khomentovskii, *et al.*, 1985). This supposition was a result of an analysis of isotope datings, distribution of phytolites and microfossils and indirect historical geological constructions.

The valid isotope datings referring to the Kerpyl' Group may be divided into two groups. On the one hand, it is old K-Ar datings of mineralogically unstudied glauconite from the Totta Formation in the middle reaches of the Maya River. They form a number of figures decreasing upsection from 1170 to 970 Ma. Two new K-Ar datings of skolite most likely Mokue borehole analogues equal to 1115 and 1120 Ma. coincide with the previous K-Ar datings. On the other hand, it is isochronous Rb-Sr and K-Ar datings of hydromica of skolite and illite composition from naturally formed outcrops along the Bolshoi Aim River. The latter correspond to much younger ages (864 ± 4 Ma.). Researchers believe that they do not represent a period of Totta rock accumulation, but rather date the processes of their regressive catagenesis (Semikhatov *et al.*, 1989). The available data allows us to conclude that the isotope age of the Totta Formation does not exceed 1170 - 1000 Ma. The datings of the overlying Lakhanda deposits give us ground to think that the age of the Kerpyl'/Lakhanda boundary close to 1000 Ma. (Semikhatov, Serebryakov, 1983). Therefore, as it was noted in the Resolutions of the 2nd USSR Conference on General Problems of Pre-cambrian Subdivision, referring the Kerpyl' Group to the Upper Riphean bears on the accepted model of the boundary between the Middle and Upper Riphean, in which the boundary is rather provisionally dated at 1000 ± 50 million years (Semikhatov *et al.*, 1991).

Three different points draw the attention of various researchers when they try to use stromatolites to define the age of the Kerpyl' Group (Semikhatov, Serebryakov, 1983; Krylov, 1985; Khomentovskii *et al.*, 1985; Komar, 1990). Some specialists emphasize the fact that *Inzeria tjomusi* Kryl., and *Jurusania cylindrica* Kryl., beginning the standard succession of the Ural Karatavian stromatolites in the Uchur-Maya section, dominate in the upper Lakhanda (Ignikan) deposits. According to this information, they draw the Yurmatan/Karatavian boundary at the base of the Ignikan For-

mation. Others underline the appearance of the mentioned stromatolite taxons near the base of the Lakhanda Group, where they are quantitatively subordinated to other forms. Drawing from these conclusions, as well as isotope datings, these researchers designate the Yurmatan/Karatavian boundary at the base of the Lakhanda deposits. Other researchers note the coincidence of the vertical succession of several stromatolite taxons in Malgina-Tsipanda, and Lower Karatavian deposits (*Malginella malgica* Kom. et Semikh., *M. zipandica* Kom., *Cono-phyton reticulatus* Kom., and *Colonella ulakia* Kom.). They correlate the mentioned deposits using historical geological criteria to compare underlying terrigenous strata (of the Totta and Zil'merdak formations, respectively). The boundary between Jurmatan and Karatavian in this case is drawn beneath Totta deposits with no regard for certain differences in their isotope datings.

Specialists studying microfossils are not even in agreement about the age of the Kerpyl' Group. Some researchers consider the microfossils that came across in this group to be the Middle Riphean ones (Timofeev *et al.*, 1976; Golovenok, Belova, 1986), and even compared the silicified Malgina microbiota with Lower Riphean microbiotas of the MacArthur Group in Australia (Golovenok, Belova, 1986). Attaching decisive importance to the *Chuaria* finds, metaphytic algae and large filamentous microremains in the lower Kerpyl' strata, the other specialists referred the considered deposits to the Upper Riphean (Veis, 1983, 1988; Pyatiletov, 1988). We shall see below that new data on composition of Kerpyl' (Totta in the first place) microfossils agree mostly with the later point of view.

Morphological Characteristic of Kerpyl' Microfossils

At present, according to their systematic diversity and quantity, organowalled microfossils are the best known microfossils of the Kerpyl' Group. They have been discovered in 24 locations at 15 stratigraphic levels in a number of sites in the Uchur-Maya region, and are represented by 42 genera and 51 species. As for the mineralized forms, they have been observed only in two locations. There are seven species of six genera in those locations and their representative sample is described in only one location.

In our classification of the considered microfossils, we observed the rules stated in a collective monograph (*Mikrofossilii,...* 1988), deviating from it only in special instances. Thus, identifying the species of the *Leiosphaeridia* genus, we use their dimensional ranges (≤70, 70 - 300 and ≥300 μm) recommended by the Leningrad symposium (Yankauskas *et al.*, 1987) and not the wider ones (≤70, 70 - 1000 and ≥1000) accepted by the mentioned monograph. We believe that the preservation of a number of typical acritarch taxons is justified. We make reference to their excellent morphological representatives such as the *Leiosphaeridia bicrura* Jank. species (envelopes with a median split) and the *Nucellosphaeridium* Tim. genus (forms with an inner body) including: *N. nordium* (Tim.), and *N. minutum* Tim. Defining genera and species of colonial coccoid and filamentous microfossils, we more frequently use those taxons which were identified in studying mineralized forms. Classification of the latter is based on more essential and biologically significant features.

Kerpyl' microfossils described below belong to 44 genera and 54 species.[1] Representatives of all the main morphological groups of the Precambrian microworld occur among them. These are acritarchs, colonial coccoid and filamentous microremains, and formations of a complex shape. Quantitatively, members of the first two groups prevail at most of the studied levels (with the exception of biotas of silicified forms). It should be emphasized that the abundance and variety of colonial coccoid forms in the considered strata is a phenomenon hardly characteristic for biotas of organowalled forms. It may be connected with specific features of the Kerpyl' paleobasin. Let us consider the most typical morphotypes in each of the mentioned morphological groups.

Acritarchs

In our material, this group of microremains is represented only by organowalled forms and includes 11 genera and 19 species. At all stratigraphic levels, the most simply organized spheromorphites of the *Leiosphaeridia* genus are predominate (more than 80% of all observed acritarchs). The size of their envelope is from the first microns to 300 μm and its thickness is from 0.5 to 2 μm. *Chuaria*, close to *Leiosphaeridia* in structure, but much larger in size, occurs less often. Their envelope diameter is 1000 - 1100 μm in the Konder and 700 - 750 μm in Omnya subformations. Less frequent are envelopes with inner bodies of different size and shape, rounded, fine *Nucellosphaeridium*, large *Pterospermopsimorpha*, and ellipsoidal *Konderia*. Envelopes at different stages of division (the smallest form diameter is 250 μm), cenobial aggregates of *Leiosphaeridia* with the diameter of 50 to 100 - 120 μm, acritarchs with a peripheral margin-fold Simia (60 - 100 μm), coarse-tubercular *Pulvinosphaeridium* (330 μm), envelopes with singular open spine *Germinosphaera* (60 μm), and acritarchlike formations with a reniform node (50 - 60 μm) play a subordinate role in the studied sample. Most of the enumerated forms (with the exception of the smallest *Leiosphaeridia*) are traditionally considered to be the remains of phytoplanktonic eucaryotic organisms that inhabited open basins with normal salinity. However, in the Vendian of Siberia (Pyatiletov, 1985) and Ukraine (Kir'yanov, 1986) combinations of spheromorphic envelopes with ribbonlike thalli fixed to substratum (e.g., the *Vanavarataenia* Pjat. genus) were discovered. This leads us to believe that at a certain stage of the life cycle at least some of acritarchs belonged to benthic associations.

[1] The most characteristic morphotypes of Totta organowalled microfossils are shown in Plates I - III.

Colonial Coccoid Microfossils

In Kerpyl' deposits, the mentioned microremains are represented by 14 genera and 15 species. Their overwhelming majority is located in the lower Totta (Konder) strata. Here they form three clear morphologically isolated associations of organowalled forms. The simplest shapeless aggregates (colonies or their fragments), deprived of a common envelope, belong to the first association. Spherical and cluster-shaped colonies including two to three to several dozen cells (sometimes with inner bodies) also belong to this association. Their diameter within such an aggregate remains constant and changes from one aggregate to another from 4 - 5 to 30 - 45 μm. Morphological features of these forms correspond to *Sphaerophycus, Palaeoanacystis, Eomycrocystis, Myxococcoides,* and *Coniunctiophycus* genera, which, probably, correspond to deep decomposition stages (capsulata stage) of various chroococcales cyanobacteria (Hofmann, 1976; Knoll, Golubic, 1979). Such forms in Kerpyl' biotas might be planktonic elements.

The second morphological association is formed by large accumulations of coccoid cells that are usually preserved as extended films and strands. Cell diameter in such accumulations varies from 8 - 10 to 20 (rarely from 30 - 35 μm) and remains the same within one accumulation. Formations from the lower parts of the Lakhanda Group that are close in structure were described as *Ostiana microcystis* Herm. by T.N. German (*Mikrofossilii,* ... 1988; German, 1990). She considered them to be an analogue of recent chroococcales cyanobacteria *Microcystis*. They form large colonies without outer sheaths and are known mainly as planktonic organisms. Their less frequent fixed varieties have also been observed. Remains of microcystislike forms are widely known in biotas from cherts or from terrigenous beds of different Proterozoic horizons (*Nanococcus, Eomicrocystis, Palaeomicrocystis* and other genera). However, they do not form so pronounced extended thalli like *Ostiana*. German interprets the described remains as two-dimensional films, consisting of coccoid cells, sometimes including inner bodies of different size (punctata stage). However, the Totta Ostiana rather often demonstrate multiple superposition of cells. This indicates an original volumetric and multilayered structure of colonies. Therefore, we may assume some part of the considered forms to be directly related to a widely distributed Precambrian genus of benthic mat-forming *Eoentophysalis* Hofm. If this assumption is confirmed, we shall have to reconsider our concepts on a specific habitat of *Eoentophysalis* in Precambrian semi-isolated oversalted sea basins.

The third association of the Lower Totta coccoid microfossils involves their most typical representatives. These are chain aggregates, different in size or morphology, formed by one, two, or three closely pressed linear cell combinations. Dense aggregates of similar cells of regular spherical shape also belong here. Among linear aggregates, we distinguish: (1) uniserial chains of cells of three-dimensional generations of 10 - 20, 20 - 30 and 30 - 40 μm; (2) biserial chains of cells, where some cell pairs (terminal and/or medial) are replaced by one larger cell; cell sizes in pairs amount to 6 - 8, 10 - 12, 15 - 16, less often up to 20 μm, and the diameter of the replacing cell ranges from 12 to 40 μm; (3) triserial aggregates, most persistent in their morphology in which the diameter of individual cells amounts to 15 - 18 μm. Among spherical aggregates of coccoid microfossils, two generations which are present were formed by cells with the diameter of 15 - 18 and 30 - 35 μm.

Combination of chain and spherical cell aggregates of silicified slates of Banff Park of the Hector Formation (Windermir Supergroup) in northwestern Canada were described by M. Moorman (1974) and called *Sphaerocongregus variabilis* Moorman. They are the closest representatives to some recent epiphytic endosporulating *Pleurocapsa* and are considered to represent different stages of the life cycle of fossil *Pleurocapsa cyanobacteria*. Large, single (more often terminal) cells in chain aggregates must correspond to endosporangia who developed under sharply deteriorating habitat conditions.

Inspite of different sizes of Totta chain and spherical aggregates and type forms described by Moorman[2], we refer the former to the *Sphaerocongregus* genus. It should be noted that Konder chain *Sphaerocongregus* colonies demonstrate similarity with some other taxons (*Polysphaeroides, Chlorogloeopsis*, and *Zinkovioides*) known from the Upper Riphean and Vendian deposits of Siberia, South Ural, and East European Platform. The problem of a taxonomic independence of all mentioned taxons requires special consideration.

In addition to the listed forms, colonial Eogloeocapsa forms with a well expressed common rounded envelope have been observed in the Konder Subformation of the Totta Formation. Similar *Eoaphanocapsa* forms with ellipsoidal sheath have been found in the Konder and Omnya subformations (cell diameter in both colonies is 10 - 12 μm). Large (greater than 150 μm) envelopes with numerous spheroidal *Aimia* inclusions, interpreted as reproductive structures of fossil fungimyxomycets are also observed in the Konder and Omnya subformations (*Mikrofossilii,*... 1988; German, 1990). In cherts of the Malgina Formation pseudofilamentous aggregates of ellipsoidal *Cyanothrixoides* cells have been described. V.K. Golovenok and M.Yu. Belova (1986) compare them with old *Tubiella* cyanobacteria.

Filamentous Microfossils

The mentioned microremains in the Kerpyl' group are represented by both trichomes and sheaths known mainly as organowalled remains. A set of filamentous

[2] Sizes of cells in chain and spherical aggregates of *Sphaerocongregus* type forms amount to 5 - 16 and 0.1 - 2 μm respectively, and of Totta forms, - to 6 - 40 and 5 - 35 μm.

septate microremains includes only the most simply structured *Oscillatoria* and *Lyngbya*-like representatives: Totta organowalled and Malgina mineralized *Oscillatoriopsis* (four species) and organowalled forms known only in Totta deposits, *Trachytrichoides* and *Filliconstrictosus* lenticular trichomes and *Palaeolyngbya* surrounded by a sheath, belong to them. The maximum width of representatives of these genera does not exceed 25 μm.

Kerpyl' tubular aseptate microfossils (sheaths) are subdivided into three morphological associations, according to their morphology and size. (1) Singular hollow tubular smoothwalled formations 1 - 60 μm wide, that were most likely sheaths of the fossil hormogonial cyanobacteria of the *Lyngbya* and *Phormidium* types (Knoll, Golubic, 1979; Mendelson, Schopf, 1982). In the studied sample, *Archaeotrichion, Eomycetopsis, Leiotrichoides, Siphonophycus*, and *Taenitrichoides* belong to them. (2) Aggregates consisting of bundles of parallel tubular formations close to each other 1 - 2 and 4 - 5 μm wide are represented in the sample by *Eomicrocoleus* and *Polytrichoides* genera. The mentioned aggregates are considered to be the remains of multiserial sheaths and/or decomposed trichomes of hormogonial cyanobacteria, analogues of recent *Microcoleus* and *Schizothrix*. (3) Tubular remains with clearly expressed transversal ribbing that measure up to 50 - 60 μm wide. Some of them have thin sharp ribs elevated above the surface and may be referred to the *Rugosoopsis* genus. Others, referred to the *Rectia* genus, have clear dense ribs formed by a combination of the finest granules that may be a result of bacterial destruction. *Rugosoopsis* is sometimes considered to be a likely representative of organisms of animal origin (*Mikrofossilii*, ... 1988). It is unclear where to biologically classify *Rectia*.

The prevailing Precambrian fillamentous microremains are interpreted as relics of benthic associations. Some of the above-mentioned taxons are known in the biotas of cherts as active mat-formers (*Eomycetopsis*) or symbionts (*Palaeolyngbya, Oscillatoriopsis*) in mat communities (Knoll, Golubic, 1979). Both occur very rarely in the studied material. In contrast to colonial coccoid forms, they evidently did not form significant bottom cenoses.

The association of filamentous microfossils is supplemented by ellipsoidal *Torgia* with thin folds on one or both poles. Such forms are very similar to ellipsoidal *Eosynechococcus* which are represented throughout the Precambrian. The latter are considered to be a morphological analogue of recent chroococcales cyanobacteria *Synechococcus* (Horodyski, Donaldson, 1980) hormogonial cyanobacteria.

Microfossils of Complex Share

These formations in the sample are represented by both large thalli of fossil algae of an unclear systematic group (possibly green or red), disrupted to a different extent, and by fungilike formations. Konder clusterlike *Ulophyton* (more than 700 μm) and Omnya three-lobed *Majaphyton* (the size of an incomplete form exceeds 110 μm) may be referred to the former. Dark bodies of irregular shape with long spines, *Caudosphaera*, may be referred to the latter. On the Lakhanda material German (*Mikrofossilii*, ... 1988; German, 1990) interpreted *Caudosphaera* as analogues of structures occurring at vegetative germination of Mucor fungi spores. All known Kerpyl' microfossils of complex shape might be benthic organisms.

Distribution of Microfossils in the Kerpyl' Group

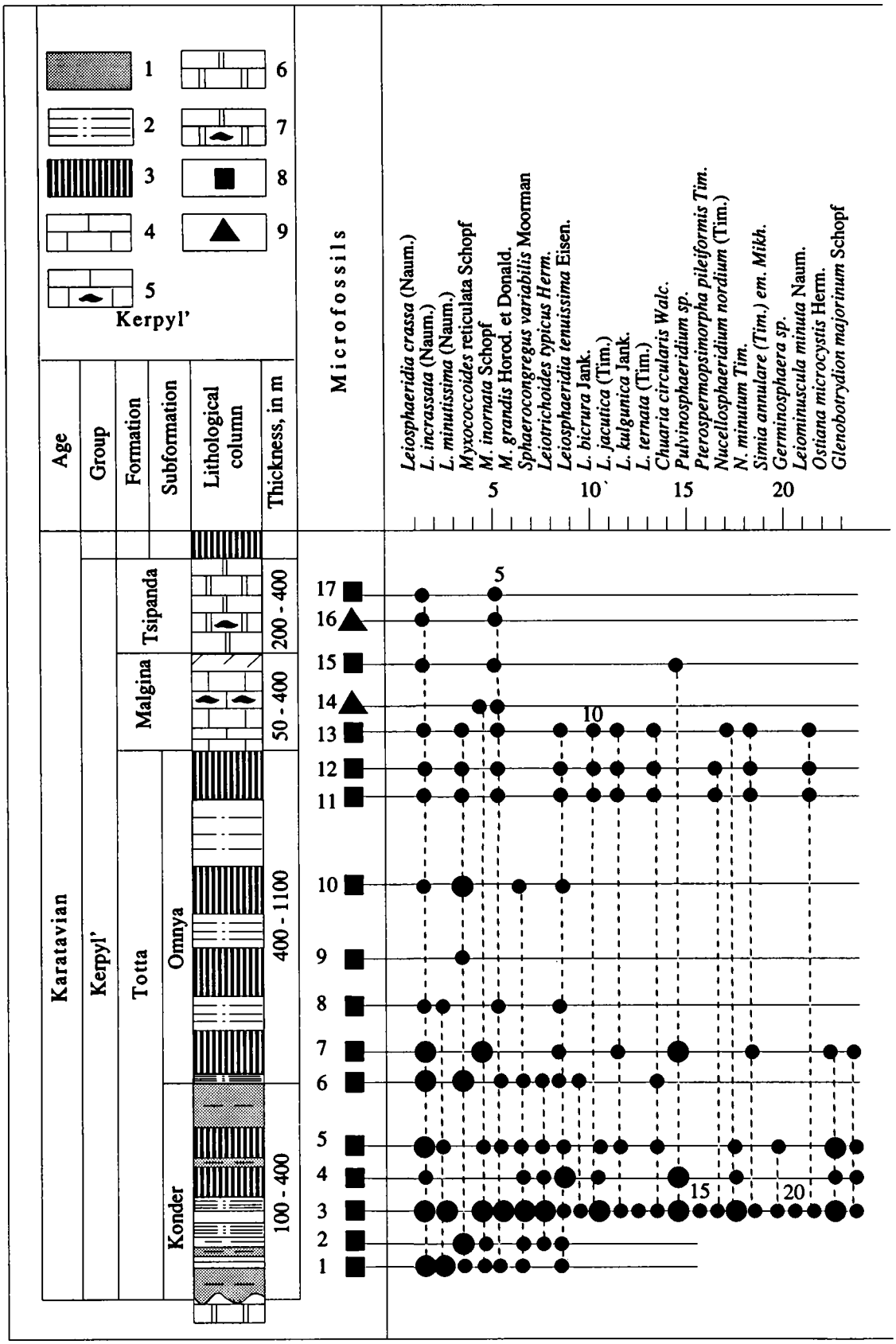
The main mass of Kerpyl' microfossils originates presently from Totta deposits, namely, from the lower Totta beds of the Bolshoi Aim River basin. In lower Totta deposits of other localities in the Uchur-Maya region and in overlying horizons of the group (especially in its carbonate part) microfossil sets, poor in quantity and systematic diversity, have been discovered. On the whole, 17 consecutive levels in the total section of the group, characterized by microfossils, may be identified. Five of them refer to the lower Totta (Konder), seven to the upper Totta (Omnya), three to Malgina, and two to Tsipanda deposits. Complete data on microfossil composition at each of these levels are shown in Fig. 2. In the following text, we shall delve into their general microphytological characteristics.

Totta Formation, Konder Subformation

The most representative samples of Konder forms have been collected on the northern slope of the Omnya uplift in the Bolshoi Aim River basin. Much less representative samples, in the north of the Yudoma-Maya basin along the Belaya River (Svetlyi and Dim Creeks).

On the slopes of the Omnya uplift, the Konder Subformation consists of three members: two thin (18 - 20 and 10 - 11 m) members of grey predominantly quartzose sandstone with subordinated interbeds of siltstone and claystone, and a thicker (78 - 80 m) member of greenish-grey and dark grey siltstone and claystone with rare interbeds of fine-grained sandstone (Semikhatov, Serebryakov, 1983). Sandstone beds in the first and third members have traces of wavemarks and rare desiccation cracks. Sometimes, they are of well expressed cross bedding. A siltstone-claystone member separating them is characterized by clearly observed thin horizontal or hammock bedding. Its rocks include numerous siderite concretions and interbeds enriched with skolite. At some levels, clear washout structures of the gutter casts type filled with fine- and medium-grained sandstones can be observed.

Microfossils occur in the first and second members. In the first, they are not numerous and are located at two levels that are 2 - 4 m from the visible base of the subformation (the right bank of the Bolshoi Aim River, 1 km below the Omnya River mouth) and 12 - 16 m from it (the Alyi Creek, the right tributary of the Bolshoi Aim River). At both levels similar impoverished sets of


1
2
3
4
5
Kerpyl'
6
7
8
9
Age
Group
Formation
Subformation
Lithological column
Thickness, in m
Microfossils
Karatavian
Kerpyl'
Tsipanda
Malgina
Totta
Omnya
Konder
200 - 400
50 - 400
400 - 1100
100 - 400
Leiosphaeridia crassa (Naum.)
L. increassata (Naum.)
L. minutissima (Naum.)
Myxococcoides reticulata Schopf
M. inornata Schopf
M. grandis Horod. et Donald.
Sphaerocongregus variabilis Moorman
Leiotrichoides typicus Herm.
Leiosphaeridia tenuissima Eisen.
L. bicrura Jank.
L. jacutica (Tim.)
L. kulgunica Jank.
L. ternata (Tim.)
Chuaria circularis Walc.
Pulvinosphaeridium sp.
Pterospermopsimorpha pileiformis Tim.
Nucellosphaeridium nordium (Tim.)
N. minutum Tim.
Simia annulare (Tim.) em. Mikh.
Germinosphaera sp.
Leiominuscula minuta Naum.
Ostiana microcystis Herm.
Glenobotrydion majorinum Schopf
5
10
15
20
17
16
15
14
13
12
11
10
9
8
7
6
5
4
3
2
1

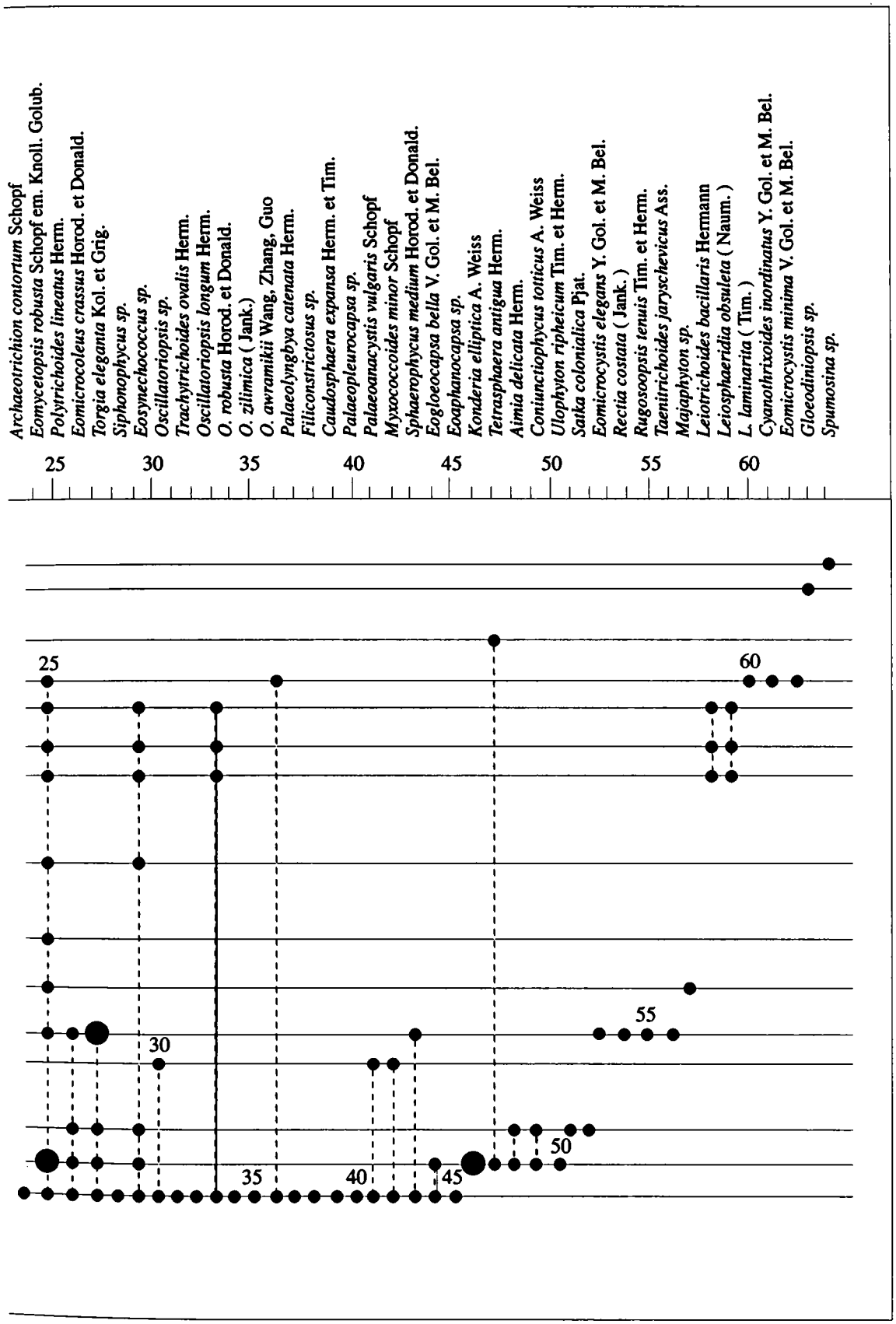


Fig. 2. *Scheme of stratigraphic distribution of microfossils in the Kerpyl' Group deposits of the Uchur-Maya region.*
1 - 7 – dominating types of the rocks of the Kerpyl' Group: 1 – sandstone; 2 – siltstone; 3 – claystone; 4 – limestone; 5 – bituminous limestone; 6 – dolomites; 7 – dolomite with lenses of early diagenetic flint nodules; 8 - 9 – levels of microfossil occurrence: 8 – organowalled; 9 – silicified.

microfossils were found, that include *Leiosphaeridia* acritarchs (envelope diameter of 20 - 114 μm), colonial coccoid *Myxococcoides* (cell diameter in colonies from 12 to 30 - 45 μm) and rare smoothwalled *Leiotrichoides* sheaths (up to 16 - 20 μm wide). Triserial chain forms of *Sphaerocongregus* are present at the second level (size of certain cells is 16 - 18 μm).

More than half of all known Kerpyl' microfossils occur in the second member. All Konder form finds described before, refer here (Veis, 1983, 1988). Their maximum amount is discovered at three consecutive levels (from the third to the fifth, counting from the base of the group), located at a distance of 2 - 10, 35 - 40 and 55 - 65 m from the member base. However, rare microremains may occur between the mentioned intervals. Microfossils of the lower level are in the outcrops of the subformation on the left bank of the Bolshoi Aim River, 10 km above the Omnya River mouth; of the middle one, on the right bank of the Bolshoi Aim river 10 - 12 km below the Omnya River mouth; of the upper one, on the left bank of the Bolshoi Aim River, 7 km above the Omnya River mouth.

A set of microfossils of the third level includes acritarchs *Leiosphaeridia* and *Chuaria* (their diameter varies from 8 - 10 to 1000 - 1100 m), envelopes with inner bodies of *Pterospermopsimorpha* and *Nucellosphaeridium* as well as more complex forms with *Germinosphaera* spine, peripheral Simia folds and coarse-tubercular *Pulvinosphaeridium*. Colonial coccoid microfossils are especially diverse here. Their cell diameter varies from 5 - 8 to 40 - 45 μm. Most simply constructed *Myxococcoides*; spheromorphic colonies of *Palaeoanacystis* and *Sphaerophycus*; *Eoaphanocapsa* and *Eogloeocapsa* with a common envelope; large *Ostiana* thalli; complex *Palaeopleurocapsa*, and numerous chain aggregates of *Sphaerocongregus* cells occur here. Biserial varieties of the latter (the least cell diameter of 8 - 10 μm) appear frequently. An important element of the considered set is the filamentous forms trichomes and sheaths. The former are represented by *Oscillatoriopsis, Filiconstrictosus, Palaeolyngbya*, and *Trahytrichoides* genera; the latter, by *Archaeotrichion, Eomycetopsis, Leiotrichoides, Siphonophycus, Eomicrocoleus,* and *Polytrichoides*. Their widths vary from 10 - 28 and 1 - 40 μm, respectively. Ellipsoidal *Torgia* and *Eosynechococcus* sheaths similar in structure occur only here. *Caudosphaera* belong to microfossils of complex shape. A set of microfossils of the fourth level is specific in composition. Against the background of a certain reduction of the diversity of present morphological groups of microfossils, a number of new forms appear here unknown to the underlying Kerpyl' deposits. Among acritarchs they are represented by large ellipsoidal *Konderia* forms with an inner body repeating outlines of the outer envelope (the size of the latter of up to 200 x 160 - 180 μm) and sporangialike *Aimia* and *Tetrasphaera* which are respectively groups and tetrads of cells in a general sheath (the outer diameter of up to 150 μm). Among colonial coccoid forms, there are cluster-shaped aggregates of spherical and subspherical *Coniunctiophycus* colonies (diameter of individual cells in a colony 15 - 16 μm) and peculiar *Eogloeocapsa* with two dense spherical aggregates of cells (cell diameter of 10 - 12 μm) in a common sheath. Infrequent *Leiosphaeridia, Nucellosphaeridium* and *Chuaria* (diameter of up to 520 μm), large *Myxococcoides* (cell size in colonies of up to 40 - 45 μm), colonies close to *Glenobotrydion* (cell diameter of up to 40 m), and rather rare *Ostiana* thalli and biserial chain aggregates *Sphaerocongregus* (diameter of the smallest cells of 20 μm) occur here. Infrequent *Eomycetopsis, Leiotrichoides* and *Siphonophycus* sheaths (from 3 - 4 to 40 μm wide) are also observed here. Of specific interest are clusterlike thalli of metaphytic *Ulophyton* which are observed only at the fourth level.

An overlying fifth level completes the sequence of Konder microfossils. The first Konder tabular acritarchs *Satka* (diameter of up to 200 - 300 μm) and cluster-shaped aggregates of small colonies *Eomicrocystis* (size of certain cells of 4 μm) make its microphytological characteristic a peculiar case. They are associated with underlying acritarchs, colonial coccoid forms, and rare sheaths (Fig. 2), whose sizes are not beyond the limits mentioned for respective taxons from an underlying level. Envelopes with spines - "buds", triserial chain aggregates (cell diameter of 16 - 18 μm), dense spherical cell accumulations (cell diameter of up to 30 - 32 μm), not numerous *Ostiana* thalli, including those at decomposition "punctata" stage, are observed here.

By their stratigraphic position in the formation, rare findings of acritarchs and sheaths observed in the middle part of the Totta Formation at compatible levels along the Svetlyi and Dim Creeks (beds 5 and 6 respectively) may be referred to the fifth level (Semikhatov, Serebryakov, 1983). They are represented by small acritarchs *Leiosphaeridia* with the diameter of 36 - 46 μm and sheaths of *Leiotrichoides* 6 - 12 μm wide.

An analysis of five sets of Konder microfossils and lithological features of enclosing rocks shows that their third set, systematically most diverse, is confined to deposits in a shallow sea basin, possibly with a changeable, but more likely with a relatively stable hydrodynamic environment. This set includes all the main morphological varieties of microfossils in which acritarchs most frequently occur. Among them are large *Chuaria* and extended thalli of colonial coccoid forms. The former were an important component of the planktonic community (together with a part of colonial coccoid and filamentous cyanobacteria), the latter, the members of a representative benthic association. *Pleurocapsa* cyanobateria and fungilike formations also occurred there.

The fourth and fifth sets of Konder microfossils located in more clayey rocks formed in much less dynamic (deep-sea?) environments are characterized by clear impoverishment of taxonomic diversity and the amount of microremains; the fifth set, is characterized by a decrease in the sizes of acritarchs as well. A similar, yet greater, tendency is demonstrated by the

first and second sets of Konder microfossils that occurred in sediments of the most shallow and highly dynamic environments. Thalli of coccoid forms are completely absent here. The number of acritarchs is reduced and their sizes are limited with the first hundred microns. It follows (under the condition that the studied samples adequately reflect quantitative correlations of Konder microfossils in their lifetime) that only a relatively narrow zone of the Konder sea basin (see characteristic of the third level) corresponded to optimum conditions of existence and/or fossilization of planktonic and benthic associations of microorganisms. We shall see below that the marked tendencies are of a general value and their analysis allows us to draw substantial conclusions about the biological connections of Precambrian microplankton.

Omnya Subformation

In deposits of this subformation organowalled microfossils occur at seven stratigraphic levels (6 - 12 levels from the base of the group). Their finds are located in central (the Maya River), western (the Omnya River), and northwestern (the Aim River) parts of the Maya basin (Fig. 1).

The sixth level is located at a four meter interval from the lower part of the Omnya Subformation. Gray thin-laminated, very sandy siltstones enclosing microremains which are exposed on the left bank of the Omnya River, 130 km from its mouth. Among the forms found here, there are a few *Leiosphaeridia* acritarchs, ascending from below, with a diameter of 10 - 60 μm; colonies of coccoid *Palaeoanacystis* and *Myxococcoides* arranged in a simple way with cell sizes of 4 - 6 to 30 - 45 μm, including rare unicatenular aggregates of smaller cells (up to 22 μm) and rare smooth-walled *Leiotrichodes* sheaths (8 - 20 μm wide) Ellipsoidal *Eosynechococcus* also occur here.

The seventh level corresponds to a 6 - 8-m member of thin intercalation of gray-green silty-clayey rocks and fine-grained sandstones. They are exposed on the left bank of the Omnya River 122 km from its mouth, 40 - 45-m stratigraphically higher than the preceding level. Here the most representative set of Omnya microfossils includes all their main morphological varieties. The most essential distinctive feature of this set is gigantic *Chuaria* acritarchs (envelope diameter of up to 650 - 750 μm), large *Leiosphaeridia* (diameter of up to 300 μm), extended *Ostiana* thalli, wide tubular forms – transversally ribbed *Rectia* and *Rugosoopsis* (up to 50 - 55 μm wide) – and smooth-walled *Taenitrichoides* (60 μm wide and more). There are numerous acritarchs *Leiosphaeridia* and *Nucellosphaeridium* (diameter up to 150 μm) passing from underlying deposits; divisible forms; colonial coccoid *Myxococcoides* and *Glenobotrydion* (cell diameter in colonies of 10 - 25 μm); *Eomycetopsis*, *Leiotrichoides*, *Polytrichoides* and *Eomicrocoleus* sheaths (1 to 16 - 18 μm wide). The list is supplemented by microfossils of a more complex structure: three-lobed *Majaphyton* thalli (more than 110 μm long) and large conic film remains (more than 250 μm in size).

Levels eight through ten are located in three 1 - 12 m siltstone-claystone members and exposed in a series of bank outcrops along the Aim River between the mouths of Chyyan and Eren Creeks. All three levels include similarly impoverished sets of taxons widely developed in over- and underlying deposits. These are *Leiosphaeridia* acritarchs (diameter of 15 - 70 μm), colonial coccoid *Myxococcoides* (cell diameter of 20 - 30 μm), and filamentous *Eomycetopsis* and *Leiotrichoides* (4 - 20 μm wide). Let us emphasize that all the samples studied at these three levels overflow with brown decomposed fragments of organic films, which are the remains of some benthic organisms (possibly thalli of large *Metaphyta*) decomposed and strongly reworked by heterotrophic bacteria. Original morphology of these remains is not accessible to restoration.

The eleventh and twelfth levels complete the succession of Omnya microfossils. They belong to a 100 m interval of the subformation section adjacent to its roof. Respective finds described by V.G. Pyatiletov (1988) are located on the right bank of the Maya River opposite the Maimakan Creek and on the right bank of the Aim River near the mouth of the Eren Creek. In both cases, microfossils originate from thin interbeds of dark greenish-gray siltstones and claystones close in lithology and alternating with their red colored differences and fine-grained sandstones. The composition of the microfossils at each of the levels is similar and includes 14 species referring to 9 genera of acritarchs, colonial coccoid microfossils and filamentous forms (Fig. 2). The diameter of acritarch envelopes here ranges from a few microns to 150 - 160 μm, and that of cells in colonies, up to a few tens of microns. The width of trichomes is more than 20 μm.

The considered Omnya biotas manifest the same taxonomic diversity dependence and dimension parameters on lithological features of the enclosing rocks that were noted above for Konder microfossils. The most diverse systematically and quantitatively rich seventh set of Omnya microfossils includes gigantic acritarchs, numerous but disconnected colonial coccoid and filamentous forms (planktonic components of the set), and extended thalli (its benthic component). Both categories of microremains are confined to the member which is lithologically closest to the one including the richest (third) set of Konder microfossils. This member in Omnya Subformation is composed of thin (usually less than a few cm) alternation of grey and dark-grey clayey siltstones and grey fine-grained sandstones. The latter, like their Konder analogues, have traces of wavemarks on bedding surfaces and semidissolved glauconite globuli. They vary in thickness along the strike of each bed and were formed in a shallow, moderately dynamic basin. Much poorer sets of Omnya microfossils, especially in samples overfilled with brown decomposed organic films, belong to monotonous unclearly bedded members of claystones deprived of any structural features indicating increased hydrodynamics, and corre-

sponding to more stable and deep-sea environments in the basin.

Comparison of conclusions on preferable facies and ecological location of Konder and Omnya acritarchs with a reliable model of lateral distribution of middle Paleozoic organowalled phytoplankton in England (Richardson, 1984; Richardson *et al.*, 1990) gives unexpected results. Within the mentioned phytoplankton from Upper Silurian and Lower Devonian, deposits there are two prevailing morphological associations. On the one hand, there are acritarchs (mainly dinoflagellate cysts). On the other hand, there are leiospheres and tasmanites (fossil analogues of phycomats of recent prasinophyte chlorophytes (Richardson 1984; Tappan, 1980). Both demonstrate their own tendencies of distribution in deposits of different facies. Acritarchs are preferably located in moderately deep-sea siltstone-claystone shelf deposits formed in zones with relatively quiet hydrodynamics. Phycomatlike formations gravitate toward either extremely shallow coastal sandstone-siltstone layers or deep-sea black shales (Tappan 1971; Cramer, Dietz, 1972). Comparison of the main features of facies adherence of Riphean Totta and Silurian-Devonian phytoplankton shows Totta acritarchs (with quantitative prevalence of *Leiosphaeridia*) to be distributed similarly to the Middle Paleozoic acritarchs. In this respect, they differ from Middle Paleozoic leiosphaeres. It may be a reflection of close biological connections of Precambrian *Leiosphaeridia* and Paleozoic acritarchs, and of the absence of such closeness between Riphean *Leiosphaeridia* and Paleozoic leiosphalres. This conclusion contradicts the earlier one on the closeness of these leiosphaeres (Zindgren, 1982).

Malgina Formation

Mainly carbonate variegated Malgina strata enclose three levels of microfossil locations (13 - 15 for the whole group). Their organowalled representatives occur at the lower and upper level, the mineralized, on the middle one.

The lower Malgina (thirteenth) level is represented by two locations (the right bank of the Maya River near the mouth of the Khaakhar Creek and the right bank of the Aim River near the Eren Creek) in the lower 20 - 30 m of the formation (Pyatiletov, 1988). Microfossils were found in thin interbeds of greenish-grey clayey-limestones occurring among their grey-colored, almost purely carbonate, varieties. Taxonomic composition and size characteristics of microfossils found here (8 genera and 14 species) are similar to those of two last sets from the underlying Totta Formation (Fig. 2).

Middle Malgina (fourteenth) level corresponds to a locality of mineralized microfossils on the right bank of the Ingili River, along the Aragastakh Creek (Golovenok, Belova, 1986). Microfossils occur here in black cherts enclosed in strata of greenish-grey limestones 100 m from the base of the formation. One of the most important elements of this set is various colonial coccoid microremains (more than half of all forms). Quantitatively, cluster-shaped accumulations of the smallest *Eomicrocystis* cells dominate among them (cell diameter in colonies varies from 0.5 to 4 - 5 μm). *Myxococcoides* occur more seldom (cell diameter is 10 - 12 μm) and a peculiar pseudofilamentous colony of ellipsoidal *Cyanothrixoides* cells is unique (cell size is 4.5 - 6.0 × 3.0 - 3.3 μm).

Rare filamentous microfossils, *Eomycetopsis* sheaths (2.0 - 3.5 μm wide) and *Oscillatoriopsis* trichomes (up to 3 - 5 μm wide), are also observed here. Some of trichomes have a sheath (up to 2 μm wide) and therefore should be referred rather to as *Palaeolyngbya* (Mendelson, Schopf, 1982).

A set of microfossils of the upper Malgina (fifteenth) level includes infrequent, highly decomposed acritarchs and colonial coccoid forms from bituminous limestones composing upper beds of the Malgina Formation along the Maya River (Talakhaia, Sygyraia and other cliffs). *Leiosphaeridia, Tetrasphaera* and *Chuaria* were defined among the former (maximum size of envelope fragments is up to 300 - 350 μm) and *Myxococcoides*, among the latter (cell diameter of 15 - 20 μm). It is extremely difficult to identifythese forms in preparations due to the abundance of orange-brown shapeless organic films.

In the Tsipanda Formation, we may provisionally identify two levels of microfossils, the sixteenth and seventeenth in respect to the entire Kerpyl' Group. Singular mineralized *Gloeodiniopsis* observed by P.N. Kolosov (1982) in cherts from the lower part of the Tsipanda Formation on the Chukhonoi River refer to the lower level. Acritarchs from Tsipanda dolomites in coastal outcrops along the Maya River refer to the upper one (Timofeev *et al.*, 1976). With due regard for the revision of spheromorphite classification, only simply constructed *Leiosphaeridia* (*Protosphaeridium*) and pitted-porous *Spumosina* may be identified in the sample collected by B.V. Timofeev (1976).

All organowalled and mineralized microfossils observed in Malgina and Tsipanda formations are represented almost exclusively by planktonic forms (colonial coccoid and filamentous microremains do not form extended thalli here). These remains do not reflect the diversity of middle and late Kerpyl' biota during its lifetime. The presence of abundant bottom cenoses in the Malgina basin is proved by a wide distribution of stratified formations of a *Malginella* type (produced by cyano-bacterial communities in carbonate layers and numerous large films, decomposed metaphytes). In the Tsipanda Formation it is proved by an abundance of morphologically expressed stromatolite formations. The inadequacy of the composition of a studied sample of microremains to paleobiota composition may be explained by a negative effect on the alkalic medium intrinsic to carbonate layers on the preservation of microorganisms (Veis, Semikhatov, 1989).

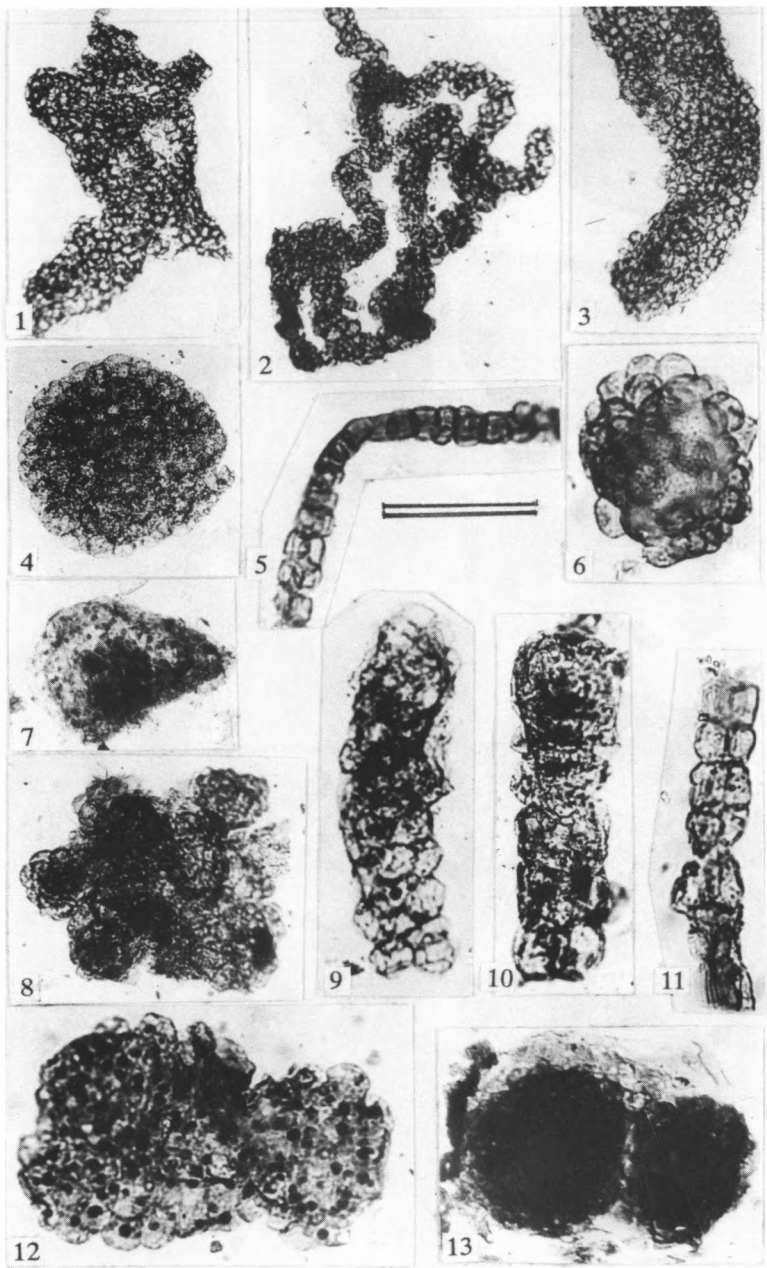


Plate I. *Microfossils of Totta Formation (Konder Subformation) of the Upper Riphean Kerpyl Group. The northern slope in the Omnya uplift of the Uchur-Maya region (East Siberia). All the presented samples are kept in Geological Institute, RAN, Moscow. 1 - 3, 5, 6, 9 - 11 – the third level of microfossil occurrence (the left bank of the Bolshoi Aim River, 10 km upstream of the Omnya River mouth). 13 – the fourth level of microfossil occurrence (the right bank of the Bolshoi Aim River, 10 - 12 km downstream of the Omnya River mouth). 4, 7, 8, 12 – the fifth level of microfossil occurrence (the left bank of the Bolshoi Aim River, 7 km upstream of the Omnya River mouth). The ruler scale size: 1, 3, 4 - 150 μm; 2 - 200 μm; 5, 6, 9 - 12 - 50 μm; 7, 8, 13 - 100 μm. 1-3 – Ostiana microcystis Hermann; thalli and cords consisting of many tens of coccoid cells; 1 – specimen no. 4022-2; 2 – specimen no. 40231-4; 3 – sample no. 40231-2. 4 – Sphaerocongregus variabilis Moorman; development stage characterized by formation of dense spherical accumulations of coccoid cells; specimen no. 4123-35.*
5, 9 - 11 – Sphaerocongregus variabilis Moorman; development stage characterized by formation of chain linear aggregates of 1 - 3 longitudinal series of cells; 5 – specimen no. 4023-10; 9 – specimen no. 4021-13; 10 specimen no. 4021-15; 11 – specimen no. 4021-18.6 – Myxococcoides inornata Schopf; specimen no. 4000-37. 7, 12 – Glenobotrydion majorinum Schopf; 7 – specimen no. 4123-41; 12 – specimen no. 4123-31. 8 – Conjunctiophycus totticus A. Veis; aggregate of subspherical colonies of coccoid cells; holotype, specimen no. 4123-62. 13 – Eogloeocapsa bella V. Golovenok et M. Belova; subspherical aggregates of coccoid cells in a common colonial sheath; sample no. 4324-44.

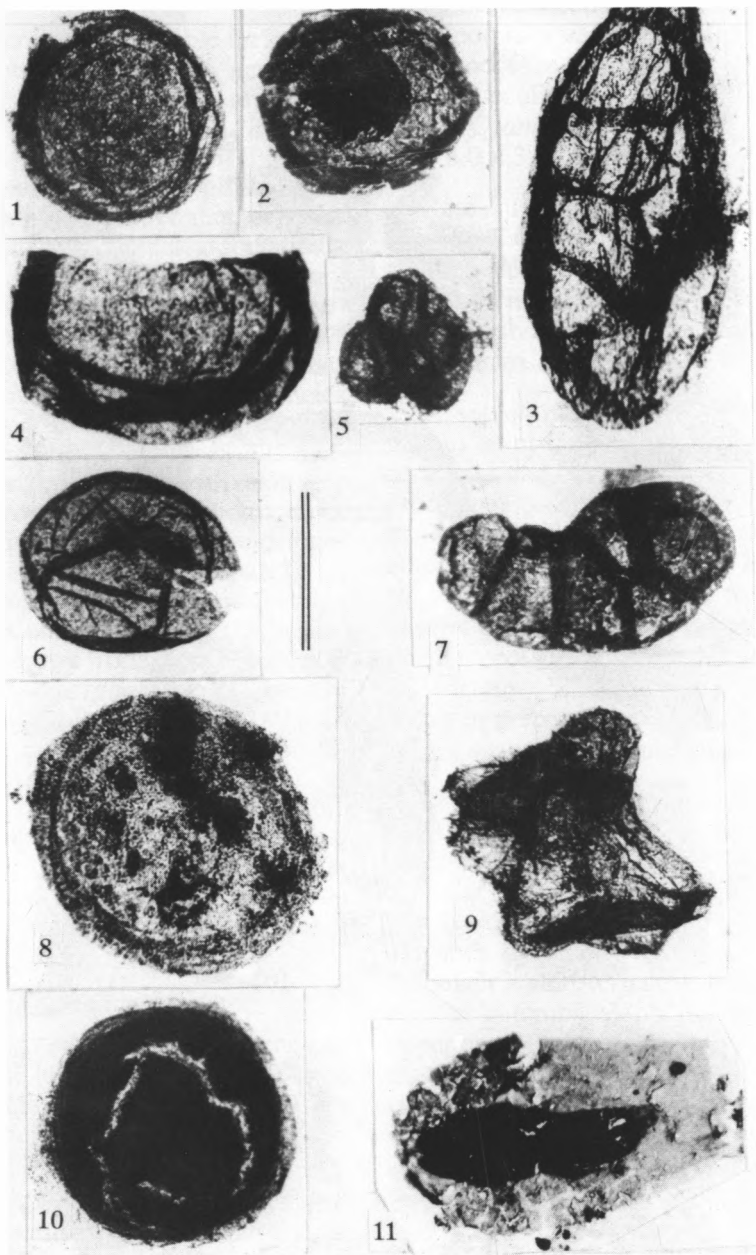


Plate II. *Microfossils of Totta Formation (Konder and Omnya Subformations) of the Upper Riphean Kerpyl' Group. Southeastern part of the Maya basin and the northern slope of the Omnya uplift in the Uchur-Maya region (East Siberia). All the presented samples are kept in Geological Institute, RAN, Moscow.*

1, 3, 4, 6 - 10 – Konder Subformation, the third level of microfossil occurrence (the Bolshoi Aim River,10 km upstream of the Omnya River mouth). 11 – Konder Subformation, the fourth level of microfossil occurrence (the right bank of the Bolshoi Aim River, 10 - 12 km downstream of the Omnya River mouth). 2 – Konder Subformation, the fifth level of microfossil occurrence (the left bank of the Bolshoi Aim River, 7 km upstream of the Omnya River mouth). 5 - Omnya Subformation, the seventh level of microfossil occurrence (the left bank of the Omnya River, 122 km upstream of the mouth). The scale ruler size: 1, 4 – 500 μm; 2 - 50 μm; 3 - 150 μm; 5, 6 - 400 μm; 7 - 25 μm; 8 - 10 - 200 μm; 11 - 110 μm. 1, 3, 4, 8 – Chuaria circularis Walcott; dense large envelopes with prevailing ring-shaped folds; 1 – specimen no. 40231-3; 3 specimen no. 40221-7; 4 – specimen no. 40231-1; 8 – specimen no. 40231-40. 2 – Nucellosphaeridium nordium (Tim); specimen no. 4123-73. 5 – Aggregate of dividing envelopes ex. gr. Leiosphaeridia; sample no. 8941-3. 6 – Leiosphaeridia bicrura Jank.; envelope with well pronounced median split; specimen no. 89403-4. 7 – Palaeopleurocapsa sp.; multicell asymmetric form; specimen no. 4023-34. 9 – Pulvinosphaeridium sp.; specimen no. 40231-6. 10 – Large dense disrupted envelope in a thin film edging; specimen no. 40221-05. 11 – Konderia elliptica A. Weiss; large ellipsoid form with an inner body repeating outlines of the outer envelope; sample no. 4324-71.

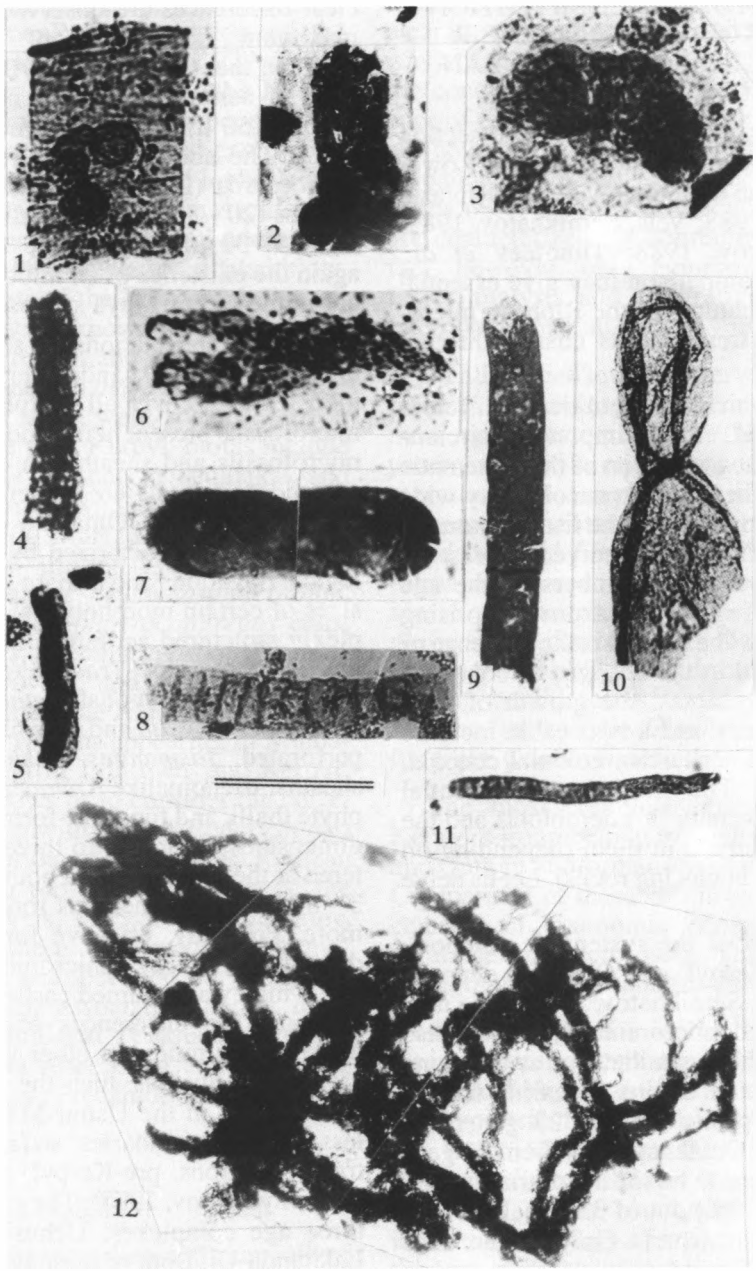


Plate III. *Microfossils of Totta Formation (Konder and Omnya Subformations) of the Upper Riphean Kerpyl' Group. South-eastern part of the Maya basin and the northern slope of the Omnya uplift of the Uchur-Maya region (East Siberia). All the presented samples are kept in Geological Institute, RAN, Moscow.*

9 – Konder Subformation, the first level of microfossil occurrence (the right bank of the Bolshoi Aim River, 1 km downstream of the Omnya River mouth). 4, 7, 8, 10 – Konder Subformation, the third level of microfossil occurrence (the left bank of the Bolshoi Aim River, 10 km upstream of the Omnya River mouth). 2, 12 – Konder Subformation, the fourth level of microfossil occurrence (the right bank of the Bolshoi Aim River, 10 - 12 km downstream of the Omnya River mouth). 5 – Konder Subformation, the fifth level of microfossil occurrence (the left bank of the Bolshoi Aim River, 7 km above the Omny River mouth). 1, 3, 6 – Omnya Subformation, the seventh level of microfossil occurrence (the left bank of the Omnya River, 122 km from the mouth). 11 – Omnya Subformation, the eighth level of microfossil occurrence (the left bank of the Aim River, downstream of the mouth of the Chyyan Creek). The scale ruler size: 1, 3, 5, 6, 9 - 11 – 60 μm; 2 - 120 μm; 4 - 40 μm; 7 - 15 μm; 8 - 85 μm; 12 - 200 μm. 1 – Rectia costata Jank.; a fragment of a wide wrinkled-ribbed sheath; specimen no. 8941-20. 2 – Konderia elliptica A. Weiss; holotype, specimen no. 4324-42. 3 – Eoaphanocapsa sp.; ellipsoidal aggregate of cells in a common colonial sheath; specimen no. 8941-23. 4 – Oscillatoriopsis sp.; specimen no. 89401-1. 5 – Elongated cylindrical form with an insignificant restriction before an increased terminal part; specimen no. 89409-7. 6 – Majaphyton sp.; strongly dissected three-lobed thallus; specimen no. 8941-23. 7 – Ellipsoidal envelope with well expressed folds near the poles close to Torgia eleganta Kolosov et Grigor. specimen no. 4023-4. 8 – Oscillatoriopsis robusta Horod. et Donald. specimen no. 4026-4. 9 – A fragment of a dense Siphonophycus sp. sheath witha thick wall; specimen no. 8952-9. 10 – Ellipsoidal envelopes at the stage of division; specimen no. 40231-12. 11 – Leiotriochoides bacillaris Hermann; specimen no. 8982-6. 12 – Ulophyton ripheicum Tim. et Hermann; a large strongly dissected branching thallus; specimen no. 4324-63.

ANALYSIS OF SYSTEMATIC COMPOSITION OF KERPYL' MICROBIOTAS AND THEIR POSITION IN SUCCESSIONS OF RIPHEAN MICROFOSSILS

Kerpyl' microbiotes are present in a single section between most representative and well studied Lower Riphean Omakhta and the deliberately Upper Riphean Lakhanda biotas (Veis, 1988; Veis, Semikhatov, 1989; German, 1990; Pyatiletov, 1988; Timofeev *et al.*, 1976). Therefore, their comparison may give essential information on general features of the Riphean microworld and on the main trends of its change through time.

Comparison of these microcommunities of different ages immediately reveals three important circumstances. The first is a great succession of the systematic composition inherent in them: the forms of a very wide vertical distribution predominate in the list of taxons of all three communities. They are observed at least all over the Riphean: the younger members of the age series including a prevailing part of taxons comprising the core of the older ones. The second is the presence of a clearly manifested trend in the changes of composition of consecutive microbiotas: the growth of their general taxonomic diversity and a noticeable increase in the maximum sizes of acritarchs, colonial coccoid, and filamentous forms. The third is the essential changes of systematic diversity of microbiotas and the sizes of microfossils observed in them (depending on lithofacies features of the enclosing rocks). Let us delve deeper into this matter.

Significant succession of the systematic composition of microfossils in Kerpyl' and Omakhta deposits was noticed earlier (Veis, Semikhatov, 1989). The new data obtained on Kerpyl' microremains make these common features even more similar, but at the same time increase the number of taxons appearing for the first time in Kerpyl' deposits. Out of 25 genera of microfossils found in the Omakhta Formation, 22 genera (more than 80%) appear to be similar to the Kerpyl' ones and 18 (more than 50%) out of 32 Omakhta species also have Kerpyl' counterparts. Only acritarchs of three Omakhta genera, *Eomarginata, Stratimorphis*, and *Polycavita* – are unknown in Kerpyl' deposits. The last two were mentioned earlier as typical for Vendian deposits of East European Platform (*Mikrofossilii...*, 1989). However this does not mean that Omnya and representative Kerpyl' microbiotas are indistinguishable. This paper has shown that the Kerpyl' biotas have much more diverse taxonomic composition (more than 40 genera and 50 species) than as was earlier thought (Veis, 1983, 1988; Pyatiletov, 1988). Kerpyl' biotas include a great number of new morphotypes. They are complex *Pulvinosphaeridium* acritarchs, *Sphaeroconcgregus* chain aggregates, multiserial filamentous *Polytrichoides* and *Eomicrocoleus*, wide tubular remains, including smooth-walled *Taenitrichoides*, wrinkled-ribbed *Rectia* and *Rugosoopsis*, rosary-shaped *Trachytrichoides*, thalli of *Ulophyton* and *Majahyton*, and fungilike *Caudosphaera*. Even more clear differences are observed in a comparison of the maximum dimensions of morphotypes enclosed between the two. In Kerpyl' deposits, the maximal sizes of acritarchs increase nearly by an order (1100 against 150 μm), of other microremains, two to three times. The above statistics refer to colonial coccoid microfossils (20 - 22 and 40 - 45 μm respectively), sheaths (20 - 22 and 50 - 60 μm), and trichomes (8 - 10 and 25 - 30 μm). Thus, the new data confirms once again the existence of an important pre-Kerpyl' boundary in the Uchur-Maya succession of microfossils.

Similar interrelations are shown by a comparison of the Kerpyl and Lakhanda (Neryuen) microbiotas. Common features cover all morphological associations of microfossils present here – most of all, colonial coccoid microfossils and sheaths. In compared biotae, practically identical sets are present. Their taxon cores are already present in Omakhta deposits. The Lakhanda biota is also characterized by quite essential morphological novations and a sharp increase of the maximum sizes of certain morphotypes. The appearance of complexly structured acritarchs must be noted here. They are acantomorphic *Trachyhystrihosphaera* and *Baltisphaeridium*, polygonal *Octaedrixium* and *Polyedrixium*, oval *Leiovalia* and *Navifusa*, lophostriatic *Valeria*, perforated, *Tasmanites*, *Lakhandinia* and *Tortunema* sheaths, trichomelike *Arctacellularia*, numerous metaphyte thalli, and fungilike forms. As for their maximum dimensions, their two to threefold increase was registered at the pre-Lakhanda boundary: acritarchs of up to 3 - 5 μm and filamentous forms of up to 100 μm and more. Presently, we have forwarded a much greater diversity of Kerpyl' microbiotas in comparison with that which was assumed earlier, it does not decrease in principle the differences of Kerpyl' and Lakhanda microcommunities. In other words, new data confirm earlier conclusions that the succession of Riphean microfossils in the Uchur-Maya region are cut by no less than two boundaries[3] so far by their most important transformations: pre-Kerpyl' and pre-Lakhanda (Veis, 1988; Pyatiletov, 1988). They divide this sequence into three age complexes: Uchur-Aimchan, Kerpyl', and Lakhanda-Ui. Both of these boundaries are of essential biostratigraphic value and are easily recognized in a type section of the Riphean in the South Urals.

New data on the composition of the Kerpyl' association of microfossils simplifies its comparison with associations of the Riphean type section in the South Urals. One of the writers (Veis, 1983, 1988) and later Pyatiletov (1988) correlated Kerpyl' microfossils and deposits enclosing them with the pre-Bederysh beds of the Zil'merdak Formation, which is a basal subdivision of the Karatavian Group (a type of the Upper Riphean). New data (Kozlova, 1987; Yankauskas *et al.*, 1989;

[3] These boundaries can be clearly revealed only in comparison with the most representative varieties of considered biotas located in platform sections of the region. If we compare their impoverished representatives (the north of the Yudoma-Maya basin), there are no essential differences (except quantitative) of microfossil associations of different ages fixed.

Veis *et al.*, 1990) were obtained recently on microfossils of the upper (Tyl'men') Subformation of the Avzyan Formation directly underlying Zil'merdak deposits, and being a part of the type of the Middle Riphean. Within the Tyul'men' biota a number of forms was found, which were earlier considered exclusively part of the Upper Riphean. They are small lophostriatic acritarchs and fillamentous remains: chain *Arctacellularia*, trichomes *Calipthotrix* and *Trachy* trichoides. These forms penetrate to the so-called Muldakai biota, belonging to the upper parts of the Zil'merdak Formation of the stratotype. They are known in the Lakhanda biota of the Uchur-Maya region and are usually compared with the Muldakai one. All these forms, however, with the exception of the latter, were not observed in Kerpyl' deposits. For their part, Kerpyl' biotas include a great number of forms present in Muldakai and Lakhanda biotas, but are unknown in Tyul'men' and its predecessors in the South Urals. Gigantic *Chuaria* (1100 against 200 - 250 μm of Tyul'men' *Leiosphaeridia*), coarse-tubercular *Pulvinosphaeridium*, chain aggregates of coccoid *Sphaerocongregus* cells, filamentous *Polytrichoides*, *Rectia*, *Rugosoopsis*, metaphytic *Ulophyton*, *Majaphyton*, and others belong to these forms. The most remarkable Karatavian morphotypes, – acantomorphic acritarchs appearing in Muldakai microbiota – are absent in Tyul'men' and Kerpyl' layers. The above-said allows us to consider the correlation of the Kerpyl' Group with the pre-Bederysh part of the Zil'merdak formation of the Upper Riphean (Karatavian) type section proposed earlier to be the most substantiated one (Veis, 1983, 1988; Pyatiletov, 1988). In the light of these constructions Kerpyl' microbiotas are a natural link between the preceding Tyul'men' (Upper Avzyan) association and a more complex and diverse Muldakai (Lakhanda) microcommunity which follows after them and continues the general tendencies of Precambrian microbiota development.

CONCLUSION

Thus, microbiotas of organowalled and mineralized microfossils of the Kerpyl' Group include a wide sample of fossil microorganisms. It consists of approximately equal groups of representatives of all their main Precambrian morphological and ecological associations. This sample includes 44 genera and 54 species of microremains, that in certain combinations (sets of forms) belong to 17 levels of different ages. Twelve of these levels belong to the terrigenous Totta, three, to the limestone Malgina, and two to the dolomite Tsipanda formations.

An analysis of sets of planktonic and benthic microfossils, different in composition in a polyfacies complex of the Kerpyl' Group, showed the richest of these sets (the third and seventh) to occur in deposits of moderately dynamic and relatively shallow sea shelf environments. Sets of sediments of more shallow and highly dynamic environments (the first and second sets) as well as those of deep-sea and stable zones (the fourth to sixth and eighth to twelfth levels) demonstrate clear impoverishment (though manifested in a different way) of their systematic composition. Unfortunately, the conclusions made about dominating facies adherence of relatively rich and poor sets of microfossils can not be verified about the materials of the Malgina and Tsipanda Formations, for the sets of organowalled forms observed here are reduced, and silicified biotas are characterized by nonrepresentative samples. This fact, however, only emphasizes the importance of facies analysis and biostratigraphic interpretation of microphytologicial data.

At all studied levels, observed microremains may be referred to planktonic and benthic associations. In the most representative sets (the third and seventh) both are present nearly in equal proportions. Among the most representative morphotypes of planktonic forms, we may name acritarchs *Chuaria, Pulvinophaeridium, Konderia*, and *Simia*; numerous colonies of coccoid forms *Myxococcoides, Sphaerophycus, Glenobotridion, Eogloeocapsa, Eophanocapsa*, and *Coniunctiophycus*; and disconnected filamentous *Eomycetopsis, Leiotrichoides, Siphonophycus*, and *Oscillatoriopsis*. Among benthicforms we may name extended thalli *Ostiana*, metaphytic *Ulophyton* and *Majaphyton*, and fungilike *Caudosphaera*.

The most complete sets of microfossils of the Kerpyl' Group belong to the lower parts of the Konder and Omnya Subformations of its basal Totta Formation. Comparison of these sets does not reveal any directed changes of taxonomic composition or dimensional parameters of microfossils. Therefore, we do not attach stratigraphic importance to the differences in these sets.

A biostratigraphic aspect of studies of Kerpyl' biotas is of specific interest in light of the discussion on the position of Yurmatan/Karatavian boundary in the Riphean hypostratotype of Siberia. Comparative analysis of the Uchur-Maya sequence of microfossils (with an account of new data on Kerpyl' microremains) confirmed an earlier conclusion on the presence of an important boundary of interregional value at the base of this group defined by a sharp widening of systematic composition of microfossils and an increase of their maximum dimensions. Comparison of the Kerpyl' association of microfossils of the Riphean hypostratotype in Siberia with biotas of its type section in the South Urals allows us to confidently consider Kerpyl' microfossils as a natural link between the Tyul'men' association belonging to the upper parts of the Middle Riphean Avzyan Formation and an association known from the upper part of the Zil'merdak Formation which itself began a type section of the Upper Riphean.

REFERENCES

Cramer, F.H. and Dietz, M. C. R., North American Silurian Palynofacies and Their Arrangement Acritarchs, *Palaeontographica Abt. B*, 1972, vol. 138, pp. 107 - 180.

Hofmann, H., Precambrian Microflora of Belcher Islands, Canada: Significance and Systematics. *J. Paleontol.*, 1976, vol. 50, no. 6, pp. 1040 - 1073.

Horodyski, R.J. and Donaldson, J.A., Microfossils from the Middle Proterozoic Dismal Lakes Group, Arctic Canada, *Precambr. Res.*, 1980, vol. 11, no. 2, pp. 125 - 159.

German, T.N., *Organicheskii Mir Milliard Let Nazad* (Organic World a Billion Years Ago), Nauka, Leningrad, 1990.

Golovenok,V.K. and Belova, M.Yu., Riphean Mlcroflora in Cherts of the Malgina Formation of the Yudoma-Maya Basin, *Paleontol. Zhurn.*, 1986, no. 2, pp. 92 - 96.

Kir'yanov, V.V., On the Nature of Some Early Cambrian Spheromorphic Acritarchs, *Aktual'nye Voprosy Sovremennoi Paleoal'gologii*, Naukova Dumka, Kiev, 1986, pp. 40 - 45.

Khomentovskii, V.V., Shenfil', V.Yu., and Yakshi, M.S., Riphean of the Siberian Platform, *Geol. i Geofiz.*, 1985, no. 7, pp. 25 - 33.

Knoll, A., Golubic, S., Anatomy and Taphonomy of a Precambrian Algal Stromatolite, *Pre-Cambr. Res.*, 1976, vol. 10, pp. 115 - 151.

Kozlova, E.V., Microfossils of the Avzyan Formation of the Middle Riphean in the South Urals, *III Vsesoyuznyi Simpozium po Paleoal'geologii Dokembriya i Rannego Kembriya: Tezisy Dokladov* (3rd All-Union Symposium on Paleoalgology of the Precambrian and Early Cembrian), Abstracts of Papers, Karek'ski Filial AN SSSR, Petrozavodsk, 1987, pp. 46 - 47.

Kolosov, P.N., *Verkhnedokembriiskie Paleoal'gologicheskie Ostatki Sibirskoi Platformy* (Upper Precambrian Paleoalgological Remains of the Siberian Platform), Nauka, Moscow, 1982.

Komar, V.A., Stromatolites in Correlation of Riphean Reference Section in Siberia and Ural), *Izv. AN SSSR, Ser. Geol.*, 1990, no. 10, pp. 30 - 35.

Krylov, I.N., *Stromatolity v Stratigrafii Verkhnego Dokembriya: Problemy'85* (Stromatolites in Stratigraphy of Upper Precambrian: Problems'85), *Izv. AN SSSR, Ser. Geol.*, 1985, no. 11, pp. 44 - 55.

Lindgren, S., Algal Coenobia and Leiosphaeres from the Upper Riphean of the Turukhansk Region, Eastern Siberia, *Stockh. Contr. Geol.*, 1982, vol. 38, no. 3, pp. 35 - 45.

Mendelson, C. and Schopf, J.W., Proterozoic Microfossils from the Sukhaya Tunguska, Shorikha and Yudoma Formation of the Siberian Platform, *J. Paleontol.*, 1982, vol. 56, no. 1, pp. 42 - 83.

Moorman, M., Microbiota of the Late Proterozoic Hector Fm., S-W Alberta, Canada, *J. Paleontol.*, 1974, vol. 48, no. 3, pp. 524 - 539.

Mikrofossilii Dokembriya SSSR (Microfossils of Precambrian of the USSR), Nauka, Leningrad, 1988.

Pyatiletov, V.G., Yudoman (Vendian) Algae of the Western Siberian Platform, *Dokl. AN SSSR*, 1985, vol. 281, no. 4, pp. 934 - 936.

Pyatiletov, V.G., Microfossils of Late Precambrian of the Uchur-Maya Region, *Pozdnii Dokembrii i Rannii Paleozoi Sibiri. Vend i Rifei*, IGiG S0 AN SSSR, Novosibirsk, 1988, pp. 47 - 94.

Richardson, J.B., Rasul, S.M., Palynofacies in a Late Silurian Regressive Sequence in the Welsh Borderland and Walles, *J. Geol. Soc.*, London, 1990, vol. 147, pp. 675 - 686.

Richardson, J.B., Palinology, Facies, and Correlation of the Middle Paleozoic, *Trudy 27-go Mgk. Stratigrafiya, Sektsiya C.01. Doklady.* (Proc. 27th IGC. Stratigraphy, Section C. 01, Reports), vol. 1, pp. 159 - 169.

Semikhatov, M.A. and Serebryakov, S.N., *Sibirskii Gipostratotip Rifeya* (Siberian Hypostratotype of the Siberia), Nauka, Moscow, 1983.

Semikhatov, M.A., Gorokhov, I.M., Kutyavin, E.P., *et al.*, Analysis of Possibilities of Sedimentary Geochronometers with Reference to the Totta Formation of the Riphean of East Siberia, *Litologiya i Poleznye Iskop.*, 1989, no. 6, pp. 3 - 18.

Semikhatov, M.A., Shurkin, K.A., Aksenov, E.M., *et al.*, New Stratigraphic Scale of the Precambrian in the USSR, *Izv. AN SSSR, Ser. Geol.*, 1991, no. 4, pp. 3 - 13.

Timofeev, B.V., German, T.N., and Mikhailova, N.S., *Mikrofossilii Dokembriya, Kembriya i Ordovika* (Microfossils of Precambrian, Cambrian and Orodovician), Nauka, Leningrad, 1976.

Tappan, H., *The Paleobiology of Plant Protists*, W. H. Freeman Co., San Francisco, 1980.

Tappan, H. and Loeblich, A.R., Geobiologic Implications of Fossil Phytoplankton and Time-Space Distribution, *Geol. Soc. Amer. Spec. Pap.*, 1971, vol. 127, pp. 247 - 340.

Veis, A.F., On the Microphytological Characteristics of the Totta Formation of the Uchur-Maya Region, *Dokl. AN SSSR*, 1983, vol. 270, no. 3, pp. 709 - 711.

Veis, A.F., Riphean and Vendian Microfossils of Uchur-Maya and Turukhansk Regions of Siberia, *Izv. AN SSSR, Ser. Geol.*, 1988, no. 5, pp. 47 - 64.

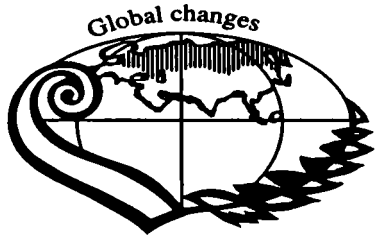
Veis, A.F. and Semikhatov, M.A., Lower Riphean Omakhta Microbiota of Eastern Siberia: Composition and Conditions of Formation, *Izv. AN SSSR, Ser. Geol.*, 1989, no. 2, pp. 36 - 55.

Veis, A.F., Kozlova, Ye.V., and Vorob'eva, N.G., Organowalled Microfossils of the Riphean Type Section South Ural, *Izv. AN SSSR, Ser. Geol.*, 1990, no. 9, pp. 20 - 36.

Yankauskas, T.V., Kozlova, E,V., and Kozlov, V.I., Tyul'men' Microbiota of the Middle Riphean of the South Urals), *Dokl. AN SSSR, 1989*, vol. 306, no. 2, pp. 419 - 421.

Yankauskas, T.V., Mikhailova, N.S., and German, T.N., Fifth All-Union Colloquium on Precambrian Microfossils in the USSR, *Izv. AN SSSR, Ser. Geol.*, 1987, no. 9, pp. 137 - 139.

Stratigraphy and Geological Correlation, Vol. 1, No. 1, 1993.Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 1, 1993, pp. 59 - 75.
Original Russian Edition Copyright © 1993 by Stratigrafiya. Geologicheskaya Korrelyatsiya, Leven.
English Translation Copyright © 1993 by Interperiodica.



Evolution of the biosphere

Main Events in Permian History of the Tethys and Fusulinids

E. Ya. Leven

Geological Institute, Russian Academy of Sciences,
Pyzhevskii Per. 7, Moscow, 109017 Russia

Manuscript received February 20, 1992

Abstract – Precise dating of Permian biostratigraphic units revealed a clear correlation between the most evident turning points of the fusulinid evolution and transgressive-regressive cycles. The two first-order cycles defined are Gzhelian-Sakmarian and Yakhtashian-Dorashamian. The lower-order cycles outlined within them are Asselian, Sakmarian, Yakhtashian-Bolorian, Kubergandian-Murgabian, Midian-Dzhulfian, and Dorashamian. Mass fusulinid extinctions are related to Late Sakmarian, Late Midian-Dzhulfian, and Late Dorashamian regressions. The first event is marked by an almost complete disappearance of the Pseudoschwagerinidae family. Larger fusulinids of Neoschwagerinida and Schwagerinida orders vanish at the second regression, and the rest of fusulinids at the third. Renewed fusulinid assemblages exist during transgressions. The appearance of *Pamirina* and *Misellina* genera, the ancestors of the Neoschwagerinida order, is related to the Yakhtashian-Bolorian transgression. The Kubergandian-Murgabian event creates *Yangchienia*, *Neofusulinella*, *Eopolydiexodina*, and *Parafusulina* genera, as well as several genera and families of the Neoschwagerinida order. Such genera as *Kahlerina*, *Reichelina*, *Codonofusiella*, *Lantschichites*, *Yabeina*, *Lepidolina*, and others make their appearance during the Midian transgression and *Palaeofusulina* and appear during the Dorashamian transgression. Links of transgressions and regressions with the eustatic fluctuations are disputable. In certain cases, these phenomena are definitely of local tectonic origin.

Fusulinids providing the main framework for Permian biostratigraphy of the Tethys ocean are already well studied. Trends and peculiarities of their evolution have been fairly well reconstructed, and principal evolutionary stages have been defined. At the same time, the development of fusulinid assemblages is treated generally without searching for relations to external events, and their influence on that evolution remains unclear because many sections are studied insufficiently. As a consequence, their inaccurate correlation does not allow us to trace essential general Tethyan processes vitally affecting sedimentation and biota evolution.

At the initial stage of the study resulting in this article, I had no intention of touching the problem mentioned above. My more modest goal assumed the revision of Permian biostratigraphy for most studied sections all over the Mediterranean - Indochina region, using my own and recently published data. Then, given the revised results, it should provide, in general, a more precise time-stratigraphic correlation. In so doing, I realized that certain events reflected in the biostratigraphy of various regions, taken at first unrelated to one another are, in fact, indicators of general steps in Tethyan evolution, the most important of which were transgressions and regressions occurring all over the Tethys area, or within its larger parts. The transgression-time curve and fusulinid evolutionary stages turned out to be well correlated, implying that time-succession of fusulinid communities and their renovations are significantly dependent on sea-level and sea-area fluctuations. Data leading to this conclusion are presented below.

Important well-expressed stages of Permian fusulinid evolution are as follows: Asselian-Sakmarian, Yakhtashian-Bolorian, Kubergandian-Murgabian, Midian, and Dzhulfian-Dorashamian (Fig. 1).

The Asselian-Sakmarian stage is characterized by the flourishing families Pseudofusulinidae, Rugosofusulinidae, and Pseudoschwagerinidae. The dominating genus of the first family was *Pseudofusulina*, which arrived at the end of the Gzhelian. Forms of the second family are abundant. Usually these are species of the *Rugosofusulina* genus, the first appearance of which is noted at the beginning of the Gzhelian. The earlier genus *Dutkevitchia* became prosperous in several Tethyan regions (Darvaz, Fergana, northern China). Not being very abundant, the third family fossils are most peculiar for the Asselian-Sakmarian fusulinid assemblages. In spite of being recently detected in Gzhelian deposits (*Paraschwagerina archaica* from Darvaz and Donbass sections, *Carbonoschwagerina minatoi* from Akiyoshi limestones, Japan, and some

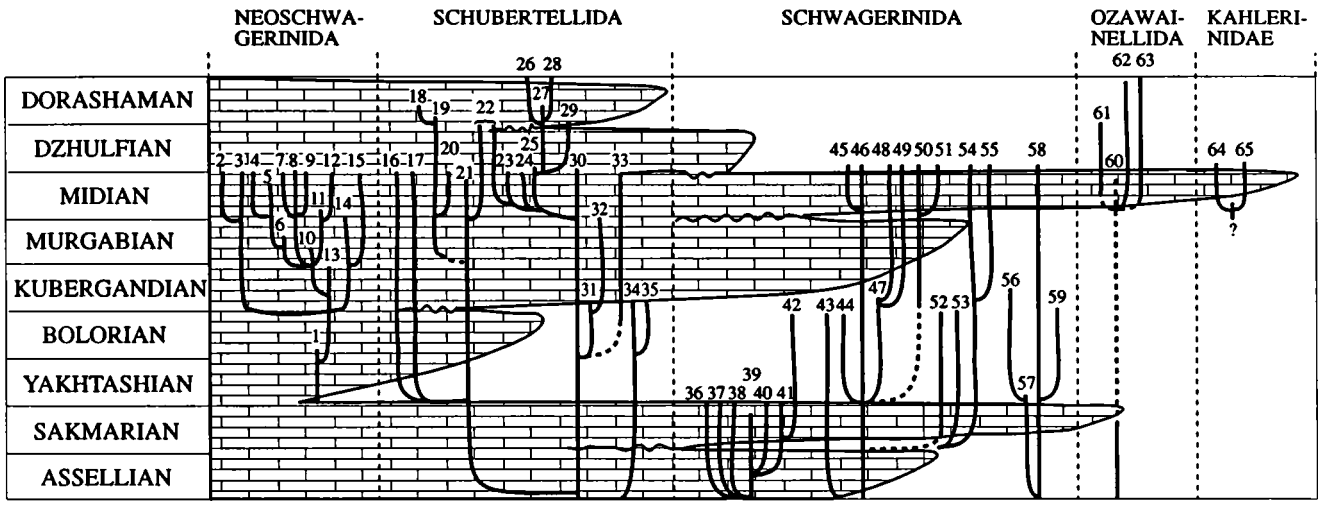


Fig. 1. *Transgressions and distribution of the most abundant fusulinid genera in the Permian section of the Tethys.*
Fusulinid genera: 1 – Pamirina, 2 – Metadoliolina, 3 – Pseudodoliolina, 4 – Sumatrina, 5 – Afghanella, 6 – Presumatrina, 7 – Yabeina, 8 – Neoschwagerina, 9 – Colania, 10 – Cancelina, 11 – Gifuella, 12 – Lepidolina, 13 – Misellina, 14 – Armenina, 15 – Verbeekina, 16 – Minojapanella, 17 – Wutuella, 18 – Gallowaiinella, 19 – Rauserella, 20 – Lantschichites, 21 – Boultonina, 22 – Codonofusiella, 23– Paradoxiella, 24 – Pseudodunbarula, 25 – Dunbarula, 26 – Palaeofusulina, 27 – Nanlingella, 28 – Parananlingella, 29 – Paradunbarula, 30 – Schubertella, 31 – Mesoschubertella, 32 – Neofusulina, 33 – Yangchienia, 34 – Biwaella, 35 – Toriyamaia, 36 – Sphaeroschwagerina, 37 – Paraschwagerina, 38 – Occidentoschwagerina, 39 – Triticites, 40 – Pseudoschwagerina, 41 – Zellia, 42 – Robustoschwagerina, 43 – Rugosochusenella, 44 – Chalaroschwagerina, 45 – Rugososchwagerina, 46 – Pseudofusulina, 47 – Praeskinnerella, 48 – Skinnerella, 49 – Parafusulina, 50 – Chusenella, 51 – Orientoschwagerina, 52 – Darvazites, 53 – Eoparafusulina, 54 – Monodiexodina, 55 – Eopolydiexodina, 56 – Laxifusulina, 57 – Dutkevitchia, 58 – Rugosofusulina, 59 – Darvazella, 60 – Ozawainella, 61 – Sichotenella, 62 – Reichelina, 63 – Parareichelina, 64 – Pseudokahlerina, 65 – Kahlerina.

quasifusulinas of the Fusulinidae family. Prosperous in the Gzhelian, Triticitedae are dying away, though in the Asselian sections of southeastern Asia they are still numerous. Schubertilidae, Ozawainellidae, and Staffellidae are of lesser importance.

The given characteristics of Asselian-Sakmarian fusulinid assemblages are not applicable to all of the Tethys. Quite a different picture is observed in its eastern Perigondwanian regions such as southern Afghanistan, southern Pamirs, Karakorum, and southern Tibet. The Asselian stage, here, is barren of fusulinids. The Sakmarian fusulinid assemblage is reduced, being represented mainly by *Pseudofusulina*, *Eoparafusulina*, and *Darvazites* forms: *Zellia* and *Robustoschwagerina* are rarely occurring. Along most southern edges of the Tethys, fusulinids are completely absent in deposits of Asselian and Sakmarian stages (Himalayas, Hindustan, Australia). Being traced in a south- and southeastern direction, this reduction of fusulinid complex, followed by their extinction, is due to climatic conditions. All of the above regions are situated at higher and colder latitudes relative to the rest of the Tethys.

In contrast to the Ural basin, where nearly all Pseudoschwagerinidae are dying out at the border of the Asselian and Sakmarian stages, the difference between the Tethyan fusulinid assemblages of the former and latter stages is not very large. It can be traced mainly at the species' level, and by the populations and diversity of certain genera.

The Yakhtashian-Bolorian stage of fusulinid evolution is marked by Pseudoschwagerinidae extinction, excluding the *Robustoschwagerina* genus. *Dutkevitchia* and most of *Rugosofusulina* forms are disappearing, as well as quasifusulinas. The specific composition of preserved genera is experiencing a rejuvenation. *Pseudofusulina* and *Darvazites* forms are dominant. Several new genera such as *Mesoschubertella*, *Toriyamaia*, *Praeskinnerella*, and *Darvazella* are beginning to appear. Appearing at the beginning of the Yakhtashian age, the *Pamirina* genus should be regarded as an ancestor for the then-prolific branch of higher fusulinids of the Neoschwagerinida order. Their first representative, the *Misellina* genus, originated in the Bolorian age, specifying the fusulinid complex of this time that is otherwise hardly distinguishable from the Yakhtashian one.

Thus, being closely related to Asselian-Sakmarian assemblages, the Yakhtashian-Bolorian complex is substantially regenerated, and already possesses some features of the younger Kubergandian-Murgabian fusulinid populations.

The differences mentioned above between North- and South-Tethyan complexes of the Asselian-Sakmarian age become less pronounced at the Yakhtashian-Bolorian time, due to the expansion of northern thermophilic forms into Perigondwanian regions of the Tethys. Fusulinid assemblages of this age from the southern regions of Afghanistan, Pamirs, and Tibet, have no significant differences either in generic, or in

specific composition relative to those from the other Tethyan regions.

The Kubergandian-Murgabian stage is characterized by a rapid differentiation of the flourishing Neoschwagerinida order. At the beginning of the Kubergandian age, two new families, Verbeekinidae and Pseudodoliolinidae, branched off of the *Misellina* genus, which first represented the order. Two other families, Neoschwagerinidae and Sumatrinidae, appear somewhat later. Each of these families consist of several genera widely present over the entire Tethys.

The appearance of several other genera is also dated by the beginning of the Kubergandian age. They belong to the following existing and new families, became abundant at the Murgabian age, and are namely: *Yangchienia, Neofusulinella, Skinnerella, Parafusulina, Eopolydiexodina, Chusenella*. The Staffellidae flourished. *Pseudofusulina* forms play a significantly decreasing role. Most other genera, such as *Darvazites, Chalaroschwagerina, Pamirina, Mesoschubertella*, and some others, guided Yakhtashian-Bolorian assemblages, totally died out just before the Kubergandian age. In general, one could argue that, at the boundary between the Bolorian and Kubergandian ages, there is the most significant and essential regeneration of the fusulinid complex in the entire history of the Permian evolution of this fossil group.

The Midian stage is closely related to the preceding one. Most of the Murgabian fusulinid genera continue to exist, having the highest level of evolution at this time. Highly advanced genera of the Neoschwagerinida order, i.e., *Yabeina, Lepidolina, Colania, Metadoliolina, Sumatrina*, are appearing. The *Chusenella* forms play a more important role among Schwagerinidae. Schubertellida has particular significance, especially the new family, Palaeofusulinidae, with *Dunbarula, Pseudodunbarula, Paradoxiella*, and *Codonofusiella* genera being noteworthy. The *Lantschichites* genus appears as a characteristic one among the Boultonidae family. A new family, Kahlerinidae, is represented by the widely appearing genera *Kahlerina* and *Pseudokahlerina*, which are found only in the Midian deposits. One should also note the appearance of *Reichellina* and Parareichellina genera, representing the aberrant Ozawainedae. These genera, as well as some forms of Palaeofusulinidae and Boultoniidae, provide a foundation for the younger Dzhulfian-Dorashamian fusulinid complex. This is why the Midian stage may be considered as transitional between Kubergandian-Murgabian and Dzhulfian-Dorashamian biotypes.

The Dzhulfian-Dorashamian stage is the final one in the history of fusulinid evolution. Nearly all fusulinids flourishing heavily at the Midian age were dying out before the beginning of the Dzhulfian age. Only certain forms of Schubertellidae, Ozawainellidae, and Staffellidae have continued to exist. They all have small shells with the outer coil volute being straight in some cases. The *Paradunbarula, Nanlingella, Parananlingella, Palaeofusulina*, and *Gallowaiinella* genera are characteristic of only Dzhulfian-Dorashamian assemblages. Several other associated genera have their origin in earlier stages. At the end of the time under consideration, all fusulinids died out.

Thus, there are four outlined episodes of the most intensive reorganization of fusulinid assemblages: the boundaries between the Sakmarian and Yakhtashian, Bolorian and Kubergandian, Murgabian and Midian, and the Midian and Dzhulfian stages. Understanding the reasons responsible for these reorganizations required the time correlation of some geological events reflected in numerous sections of various Tethyan regions. There were earlier attempts, including those of my own, to reveal such correlations. However the suggested frameworks of the correlation were insufficiently precise to obtain the desirable results. Certain general regularities, congruous with fusulinid evolutionary stages, began to be visible, due only to additional data on Tethyan regions which had been poorly studied, and after a thorough revision of all known observations.

A review of Permian Tethys deposits for each of the outlined stages of fusulinid evolution is presented below. Special attention is given to date the events resulting in hiatuses and unconformities within sedimentary sequences under revision.

ASSELIAN AND SAKMARIAN STAGES

Deposits of these ages are more widespread and represented in southeast and central Asia. Their carbonate facies are dominant in southern China. Fusulinid limestones of the Asselian and Sakmarian ages, with the same Upper Carboniferous rocks, are classified here as the Maping stage (Zhang *et al.*, 1987). As a part of the thick carbonate Backson series, they appear in Vietnam. Similar sections are observed in some regions of Japan (Akiyoshi and Atetsu stratotypes) (Nogami 1961; Ozawa *et al.*, 1991).

In the northern provinces of China, the carbonate facies are replaced by terrigenous coal-bearing deposits (the Taiyuan series). Further to the north in Inner Mongolia, there are thick chert-shaly and volcanogenic geosynclinal formations having fusulinid-bearing Asselian limestones in their composition (Zhang *et al.*, 1987). The allochthonous limestones of that age were observed in the Sulankher region of southern Mongolia (Ruzhentsev *et al.*, 1989).

In the northwestern provinces of China, the marine terrigenous and carbonate-terrigenous deposits of Asselian and Sakmarian stages are known in the Altyntag Range, west and south Kunlun Shan, and the Kelping Mountains of Kashgaria (Zhang *et al.*, 1987). They extend to the west inside the Fergana (Bensh *et al.*, 1972), and Transalai Ranges. Along the structure curve, they are detectable to the southeast, up to Darvaz and the Afghani Badakh Shan (Leven, Shcherbovich, 1978), and then to the western Kundus and Surkhob river basins of northern Afghanistan (Leven, 1971).

Characterizing the Asselian and Sakmarian deposits of the eastern Tethys regions, one can note the following:

1. Except in the northern Chinese provinces, the Asselian-Sakmarian deposits are represented by marine facies with carbonate rocks playing an essential role. The limestones are rich in fauna including fusulinids such as *Sphaeroschwagerina*, *Pseudoschwagerina*, *Zellia* and other genera of the Pseudoschwagerinidae family.

2. The Asselian and Gzhelian deposits are closely linked, displaying transitional lithology and paleontology. As a rare exception, there is unconformity at the base of the Chuanshan Formation near Nanjing (A Guide to Field Excursions in Tangshan, Nanjing 1979). Here, the Moscow limestones (Layer 42) are covered directly by carbonate breccias (Layer 43) with fusulinids of the Asselian basal zone, including *Pseudofusulina kljasmica*, *Triticites fornicatus*, *Pseudofusulina explicata* (my own determinations). In the Shanxi province of northern China, previously unmentioned unconformity seems to be present in the Taiyuan series section between the layer K_0 with *Triticites* forms of the Lower Gzhelian, and the Layer K_2 bearing the Asselian *Sphaeroschwagerina* and *Liharevites* (Rui, Hou, 1987).

When there are both Asselian and Sakmarian deposits present in the sequences, they almost always exhibit a gradual movement towards one another. However, there are data missing on the Asselian stage and the Sakmarian strata directly overlying the Carboniferous deposits. Thus, in the Kitakami Mountains of Japan, the Middle Carboniferous deposits are unconformably overlain by the Sakamotozawa series, which contains typical Sakmarian *Robustoschwagerina* and *Zellia* above the basal conglomerates and sandstones (Kanmera, Mikami, 1965). The nonevident unconformity is detectable inside the oolitic limestone sequence of the Atetsu Plateau, the lower part of which bears *Pseudoschwagerina nakasawai* typical for the Upper Carboniferous, but the upper one contains the Sakmarian *Robustoschwagerina subsphaerica* (Nogami, 1961). The pre-Sakmarian unconformity seems to be present in some rock sequences of southern China. I have come to this conclusion, revising the fusulinid collection from a section in the Guanxi province, where *Paraschwagerina* and *Zellia* begin to appear above the *Triticites*- and *Daixina*-bearing Upper Carboniferous beds. In northwestern Xingiang, the *Paraschwagerina*-bearing Sakmarian strata pseudoconformably lie over the Devonian deposits (Zhang, 1963). The Upper Carboniferous sediments in the north of southwestern Darvaz are covered directly by the ammonoid-bearing Sakmarian strata (Leven, 1981). Finally, in the Banga and Namakhab river basins of northern Afghanistan, the Sakmarian deposits were noted as resting on the Upper Carboniferous, or older strata (Leven, 1971).

Western Australia, southern Tibet, Karakorum, Hindu Kush, Himalayas, southern Pamirs, and southern Afghanistan, corresponding to peri-Gondwanan parts of the eastern Tethys, have substantially different types of the Asselian-Sakmarian deposits. Here, they are represented by terrigenous or, locally, by volcano-terrigenous formations. Thin limestone layers appear just at the top of this time range section. In the above regions, beyond the Pamirs and Afghanistan, diamictites of glacial origin were recognized in the lower parts of terrigenous formations. A faunal complex of these deposits is quite peculiar. It consists of cryophilic conulariid, bivalves (*Euridesma*, *Oriocrassatella*), brachiopod (*Stepanoviella*, *Tomiopsis*, *Taeniothaerus*), random solitary corals (*Lutvolasma*), and bryozoan. The rare ammonoids are the exact guide fossils for dating these deposits. Fusulinids appear in the upper parts of the formations, where there are limestones. The most representative collections of the last fossils are from the Bartang-Rangkul zone of the southern Pamirs. Here, they occur in the limestones of the Dangikalon Formation as a monotonous complex dominated by endemic *Pseudofusulina* forms. The *Darvazites*, *Eoparafusulina*, and rare *Zellia* species occur as well, providing the basis to classify corresponding strata as Sakmarian (Dronov, Leven, 1971). Similar fusulinid assemblages were recognized in the Chapursan Valley of Karakorum (Silva, 1965), as well as in some sections of southern Afghanistan. This last case is illustrated by the B_2-limestones of the Vardak section (Lys, Lapparent, 1971) and P_2-zone of the Tesak stratotype (Vachard, 1980).

The endemic character of peri-Gondwanan fauna, as well as the rare occurrence of orthostratigraphic groups, create difficulties for the precise definition of boundaries either between the Sakmarian and Asselian, or Asselian and Gzhelian stages. Unconformities between these stages are unknown, and the sequences seem to be continuous. However, as in the more northern regions of the Tethys, the Sakmarian deposits may sometimes (the Bartang-Rangkul zone) overlie the older strata with sharp unconformity.

The Asselian-Sakmarian sediments are less abundant in the western part of the Tethys. In many regions here, they are poorly studied, and do not differ from the underlying deposits. In the Tabas Mountains of central Iran, these sediments are recognized in the upper part of terrigenous Sardar Formation (Ruttner *et al.*, 1968). Recently, D. Bakhbani detected Sakmarian fusulinids in the section near Abadeh, where they are related to the carbonate-terrigenous Vazhnan strata. In the Elburz Mountains, the fusulinid limestones dated roughly as Asselian-Sakmarian are characteristic of the middle part of the Dorud Formation, which consist mainly of sandstones with abundant red beds, and has basal conglomerates overlying the Lower Carboniferous limestones of the Geyrud Formation Asseretto; (Kahler, 1976). The Dorud Formation is partly replaced along the strike by the terrestrial deposits. It is likely that sigillarid-bearing terrestrial sandstones from the base of the Permian sequence in the Zagros Mountains are also of the time range under consideration.

The Asselian-Sakmarian fusulinid limestones are recognized in the eastern Taurus nappes, where they are conformably underlain by the same lithofacies of the Gzhelian stage (Lys, 1986). To the west, in the Antalya region, the deposits under discussion are represented by terrestrial facies (Lys, Marcoux, 1978).

The marine, mainly carbonate, deposits of the Asselian and Sakmarian ages are widespread in the Carnic and Julian Alps, as well as in the Karawanken, and Velebit and Lika ranges, in Croatia (Kochansky-Devidé, 1959; Ramovš, Kochansky-Devidé, 1979; Kahler, 1983). F. Kahler places them into the Rattendorf and Trogkoffel stages, pointing to their conformable occurrence above the Upper Carboniferous Auernig member. Separate outcrops of the same age deposits are known also in Greece and Tunisia.

The Asselian-Sakmarian time ended in the regression that spread all over the Tethys. For southern Europe, (Kahler, 1980) relates the sea retreat to the Saalic phase of Hercynian orogeny. The uplifts took place within Hercynides of the Ural-Mongolian belt, and were reflected on the south-Chinese platform, as well as in the peri-Gondwanan Tethys regions. Everywhere, with a few exceptions, there are unconformities between the Sakmarian and overlying deposits. One can confirm their absence only in several well described sections of southern China. The marine sedimentation was probably continuous in the Darvaz. However, the post-Sakmarian uprise is notable even here by the transition from relatively deep-water flysch deposits of the Khoridzh Formation to the coarse-grained tufaceous sandstones of the Zygar Formation.

The available data enable us to state that the sea was regressing quickly, and nearly at one time all over the Tethys. Unconformities above the top of Sakmarian deposits are detectable everywhere. The Yakhtashian and Bolorian sediments, when present, are always above the unconformity surface. The only contradiction, reported by Kahler, is the fusulinid occurrence in the upper Trogkoffel (the Goggau Limestone) interpreted as an assemblage of *Pamirina darvazica-Chalaroschwagerina vulgaris* zone of the Yakhtashian stage. Apart from *Pamirina* forms, the rest of the fusulinids in this occurrence rather resemble the Sakmarian assemblages. This note also refers to "*Pseudofusulina vulgaris*", which is, in my opinion, the misdefined *Dutkevitchia* ex gr. *splendida*, the form characteristic to upper Asselian, and transferring into the Sakmarian stage. However, the occurrence of *Pamirina* forms enables the assumption that the top of Goggau Limestone is transitional to the Yakhtashian strata to be made.

YAKHTASHIAN AND BOLORIAN STAGES

Mainly carbonate, and to a lesser extent terrigenous, deposits of the age in question display gradual transitions to underlying sediments only in some areas of southern China. A discrepancy has not been observed in the Darvaz stratotype, although there, the sharp facies change mentioned above occurred at the boundary of the Sakmarian and Yakhtashian stages. For the rest of the cases, the Asselian-Sakmarian and Yakhtashian-Bolorian formations have in-between transgressive contacts with more or less evident signs of erosion and unconformity. The lack of detailed study for most sections, and a similarity of Yakhtashian and Bolorian fusulinid assemblages, occasionally do not permit us to define which of the stages is observed in an outcrop: They are often perceived as a single unit. There are certain cases when the transgressive strata are definitely of the Bolorian age. The Yakhtashian Stage is missing in such sections, or occurs as coarse-grained basal horizons.

In the Sakamotozawa-Nagaiwa area of the Kitakami Mountains (Japan), the upper unit of the Sakamotozawa Formation belongs to the Yakhtashian Stage. It has basal conglomerates, lying with the erosional contact over the lower unit which contain Sakmarian fusulinids (Kanmera, Mikami, 1965). In the nearby Setamai-Yahagi area, the Sakmarian Stage is absent in the section. At the base of the Sakamatozawa Formation there, there are fossil-free basal conglomerates and sandstones, overlying the Lower Carboniferous Nagaiwa limestone. Directly above, there are Bolorian *Misellina-bearing* Limestone, which allows us to assume the Yakhtashian age for the basal strata (Choi, Saito, 1968).

In southern China, the deposits of the age under consideration occur at the base of cherty-carbonate Chihsia Formation. As a rule, it overlaps transgressively the underlying deposits, indicated by such hiatus signs as bauxites, coal-bearing rocks, or beds with the terrestrial plant remnants, fresh-water mollusks, and ostracods.

The lower boundary of the formation is diachronous. In the most complete sections (Guizhou, Hunan, and Yunnan provinces), it corresponds to the lower boundary of the Yakhtashian Stage, and transitions to underlying Sakmarian limestones are more or less gradual here. The Yakhtashian Stage in these cases is relatively well characterized paleontologically, and Chinese scientists consider its deposits (similarly to my own interpretation) as a separate biostratigraphic unit of the stage rank (variably termed as Xianzhou, Longlin, or Houtsan) (Zhang, 1984; Xiao *et al.*, 1986; Rui *et al.*, 1987). In the sections with unconformably lying Chihsia formation, either the Yakhtashian stage, or its larger part, is missing, and the formation sequence begins with the Bolorian, or even younger Kubergandian deposits. The stratotype of the formation near Nanjing can serve as an example (Wang, 1978). At the base of the formation here, there are "Swain limestones" (Chen, 1934) with the lower strata bearing fusulinids characteristic both for Bolorian and Yakhtashian stages (*Chalaroschwagerina galloway, Darvazites ordinatus, Praeskinnerella kushmani*). However, the occurrence of *Misellina* claudiae in that assemblage precludes us from considering it older than Bolorian. Similar features were described by Sheng

et al., (1985) in sections of the Guizhou and Guangxi provinces.

The Chihsian transgression expanded northward to Chinling, where the fusulinid-bearing strata of the Yakhtashian-Bolorian age were noted. There are no data on synchronous deposits in northwestern China. They may be present in the Kunlun Shan, whose structures continue the northern Pamirs-Darvaz structures, having the same strike.

In northern Pamirs, the Yakhtashian and Bolorian stages are recognized as members of the Karakul zone sections being represented in them by thick limestone formations, including the reef facies of the rocks (Leven, 1967). They have tectonic contacts with older formations, in which there are no deposits older than the Middle Carboniferous. This indicates that Upper Carboniferous Lower Permian strata are missing in the sequence here. In the Darvaz stratotypes, the Yakhtashian and Bolorian stages are represented by thick terrigenous and carbonate (reef) strata described here as Zygar, Chelamchi, and Safetdaron formations (Leven, 1979, 1980; Leven, Shcherbovich, 1978). The upper Sakmarian and lower Bolorian deposits of this stratotype have poor faunal characteristics, and their interrelations are poorly understood, but suggested as conformable.

The Yakhtashian deposits of northern Afghanistan are studied in the Surkhob River basin. They are represented by limestones, underlain by sandstones, shales, and limestones of an unknown age, which seems to be not older than Sakmarian. There is a sharp unconformity between these rocks and the Carboniferous volcano-sedimentary strata. Somewhat farther to the south, the Bolorian fusulinids occur at the base of the carbonate-terrigenous strata, overlapping transgressively by its basal conglomerates the Carboniferous, and older metamorphic rocks (Leven, 1982).

The post-Sakmarian transgression is clearly evident in peri-Gondwanan sedimentary sections of the eastern Tethys. While, as mentioned, the Asselian-Sakmarian deposits here are predominantly terrigenous, the younger sediments are represented mainly by carbonate facies. Since the Yakhtashian Stage, the cryophilic fauna of Gondwanan type is replaced by thermophilic Tethyan assemblages. In general, the sections are poorly studied, and this sometimes precludes correctly defining the regression start-time, which, in addition, can be somewhat diverse in various regions. For example, in the Rutog area of southwestern Tibet, the first horizon of fusulinid limestones in the section contains *Pamirsina chinlingensis* and *Chalaroschwagerina tibetica*, the forms which are typical Yakhtashian, but also occur in the Bolorian Stage (Liang *et al.*, 1983). In the Shaksgam River Valley of northern Karakorum, the transgressive strata begin with the Shaksgam Formation, bearing, at the base, the upper Artinskian (Yakhtashian) corals *Verbeekiella australis* and conodonts *Merrilina* cf. *oertli* (Desio *et al.*, 1991). Somewhat higher, there are Bolorian *Chalaroschwagerina vulgaris*, *Darvazites wissi*, and *Monodiexodina karakorumensis* described by M. Reichel (1940).

The post-Sakmarian unconformity is well expressed in the southeastern Pamirs sections. In some of them (Kurteke), sandstones and shales of the Bazardara group, containing Sakmarian brachiopods, are overlain by limestones with abundant Bolorian fusulinids such as *Misellina parvicostata*, *Darvazites ordinatus*, and *Chalaroschwagerina globosa* (Leven, 1967). In other sections, the base of the transgressive series above the unconformity consists of terrigeneous rocks, bearing *Monodiexodina* similar to the forms from the corresponding parts of the Karakorum sections. At this level, and somewhat higher, there are limestone interlayers with upper Bolorian *Misellina*, *Darvazites*, and *Chalaroschwagerina* forms. Using all these data, one can argue that the Yakhtashian stage is missing in the southeastern Pamirs sections.

Similar features can be observed in many localities of southern Afghanistan. The best studied Vardak and Tesak sections are in the Logar River Valley (Lys, Lapparent, 1971; Vachard, 1980). In the first section, fusulinid limestones at the top of the Sakmarian Stage were classified by M. Lys and F. Lapparent as the Artinskian Stage, but their definition of two fusulinid genera was not very precise. My more complete fossil collection allows me to correlate them with fusulinids of the Dangika group in the Bartang-Rangkul zone of Central Pamirs. *Zellia* and *Robustoschwagerina* forms from these rocks clearly indicate their Sakmarian age. In the Vardak section, there are quartz-pebble conglomerates, passing into red sandstones, and lying with erosional contact over the B_2-limestones. Directly above them, there are clayey-carbonate deposits of the Murgabian Stage base. The sandstones are fossil-free. Their position in the section allows us to consider them as Kubergandian, or, in the lower strata, possibly Bolorian. The Yakhtashian Stage seems to be missed here again.

The B_2-limestones correspond to the rocks of the P_2 zone in the Tesak section according to the same assemblage of Sakmarian fusulinids. Calcareous breccias and oolitic and oncolite limestones of the upper P_3 zone contain *Pseudoreichellinella Darvazica*, the Yakhtashian-Bolorian species, together with Kuberghandian *Neofusulina* and *Skinnerella* forms, as reported by D. Vachard. For this part of the section, I have recognized *Darvazites ordinatus* and *Chalaroschwagerina vulgaris* in the collection of S. Karapetov. These forms can be either Yakhtashian or Bolorian, but the latter seems to be more correct because Kubergandian fusulinids occur directly above this part of the section. With this interpretation, a hiatus corresponds to the Yakhtashian Stage.

The presence of Yakhtashian-Bolorian deposits at the base of transgressive carbonate strata is also notable in the more western Tethyan regions, e.g., in Iran and Transcaucasia. In the Tabas Mountains of central Iran, the eroded terrigenous Sardar Formation is overlain by basal conglomerates of the thick limestone-dolomite Djamal Formation (Ruttner *et al.*, 1968). The lower

member of the latter is 40 - 50 m in thickness, and consists of calcareous shales with interbeds of fusulinid limestones. Described by Kahler (1974), these fusulinids belong to the typical upper Bolorian assemblage: *Misellina sp., Chalaroschwagerina vulgarisiformis, Pseudofusulina kraffti, Ps. postkraffti* etc. The Asselian-Sakmarian age of the underlying Sardar Formation is tentatively accepted, and the problem of the Yakhtashian Stage cannot be unambiguously solved. However, similarly to other described sections, one may state with confidence that this stage is missing.

The Abadeh region is the second locality of central Iran, where there are deposits of the age in question. Here, the terrigenous-carbonate Vazhnan strata, bearing Sakmarian fusulinids, are overlain by the carbonate Surmag group, which seems to begin mainly with the Kubergandian stage. However, one of the Surmag sections has, at the very base, strata with the Yakhtashian-Bolorian fusulinids: *Darvazites cf. ordinatus, Cuniculinella aff. rotunda*, and others (Taraz, 1974; Iranian-Japanese Research Group, The Permian and the Lower Triassic Systems in Abadeh Region, Central Iran, 1981).

Deposits of the Yakhtashian and Bolorian stages (the Davali Formation) are well known in the Saripap Hills section of southern Armenia. The Permian carbonate strata here overlie the Visean limestones. Their lower limestone member, 50 m thick, bears the Yakhtashian-Bolorian *Darvazites ordinatus* and *Chalaroschwagerina vulgaris* in its lower part, while the Bolorian *Misellina minor* and *M. parvicostata* are present in its upper part. Upsection, there are Kubergandian limestones with the red conglomerates at the base (Leven, 1975).

The westernmost outcrops of the deposits of the same age are known in Montenegro (Kochanky-Devidé, Milanovič, 1962). They are isolated outcrops of limestones with the Yakhtashian-Bolorian fusulinids: *Chalaroschwagerina vulgaris, Darvazites taraensis, Biwaella europae*. It is still unknown how they are positioned in a sequence, and what their relations are with older and younger units.

KUBERGANDIAN AND MURGABIAN STAGES

Deposits of these ages are much more widespread than the Yakhtashian-Bolorian sediments, thus evidencing an expansion of transgression. They are mostly represented by carbonate facies, being conformable relative to the Yakhtashian-Bolorian strata when the latter are present. In only a few sections, a kind of hiatus may be suggested due to conglomerates, which appear above the older rocks. In many sections, the deposits under discussion lie over the Sakmarian, or older formations. Their lower boundary can be diachronous, progressing from lower Kubergandian to lower Murgabian zones.

Data from Japan on the biostratigraphic interval in question are fragmentary, and do not allow a clear understanding of what is occurring at the transition between the Kubergandian-Murgabian and underlying deposits. The boundary seems to be gradual in many cases. The unconformity between the Kubergandian *Parafusulina kaerimisensis* Zone, and the underlying Bolorian *Misellina claudiae* Zone has been detected in the well known Akiyoshi section (Ota, 1977). In the other well known Kitakami section, the Kubergandian-Murgabian time coincides with a gap in sedimentation (Saito, 1968; Choi, 1973).

The Kubergandian and Murgabian deposits are widespread in Indochina, but again, their relations to underlying sediments remain unclear.

Everywhere in southern China, the deposits under discussion can be recognized in the lower part of the Chihsia Formation. When it begins with the Yakhtashian-Bolorian sediments, the transition to Kubergandian strata is gradual. Yet, in some areas, the Yakhtashian and Bolorian stages are missing in sequences, and the base of transgressive Chihsia Formation corresponds to the lower boundary of the Kubergandian Stage.

However, according to Chinese publications, this last point is not so evident. The "Chihsian stage", as defined by Chinese colleagues, begins with the *Misellina* zone. Recognizing these forms in various places at the base of the Chihsia Formation, they automatically define the formation base as the lower boundary of that zone: in other words, as the beginning of our Bolorian Stage. However, *Misellina* forms at the base of the formation belong to a highly developed species, such as *Misellina claudiae*, which is typical of transitional Bolorian-Kubergandian strata, or *M. ovalis*, the index species of the Kubergandian lower zone. That is why one may confidently state that the base of the Chihsia Formation is close to the boundary between the Kubergandian and Bolorian stages, or coincides with it.

The Kubergandian-Murgabian deposits are observed as members of geosynclinal strata in Chinling, Inner Mongolia, Altyntag, and Kunlun Shan. It is unclear how they lie over the older deposits. In the Karakul zone of northern Pamirs, along the western continuation of the Kunlun Shan structures, the lower Kubergandian zone is represented by marls, siltstones, and sandstones of the Variegated Formation. The rocks abruptly replace, upsection, the Bolorian limestones underneath. The quartz-pebble conglomerate interbeds at the base of Kubergandian strata may indicate a brief stratigraphic hiatus before their deposition. Upsection, the lower Kubergandian strata are replaced by fusulinid limestones of the upper Kubergandian and Murgabian stages (Leven, 1967).

In the Darvaz region, the transition between Bolorian and Kubergandian deposits is gradual. The latter are mainly terrigenous, though contain large bodies of algal reefs. The marine Kubergandian facies grade upsection into variegated and red terrestrial tuffs, sandstones, and conglomerates. These clearly regressive features of Kubergandian-Murgabian deposits in Darvaz exist against the background of a general transgression, and are the result of uplift movements, which took

place at the end of the Sakmarian in Tien Shan, and then extended southward. The same tendency is traceable in northern sections of northern Afghanistan (the Surkhob River Valley), where a continental regime has already begun at the beginning of the Kubergandian, or possibly even during the Bolorian. In the southern regions of northern Afghanistan (the Bamian and Bulola rivers), as well as in the Karakul zone of northern Pamirs, the Kubergandian-Murgabian sections display transgressive sequences. The Bolorian rocks dominated by sandstones and shales are replaced here by limestones of the Kubergandian and Murgabian stages (Leven, 1982).

In the peri-Gondwanan regions of eastern Tethys, deposits of this age are widespread, but poorly identified due to the limited survey. In southern Tibet, near the Rutog, the Kubergandian *Parafusulina* and *Skinnerella* forms were recognized in the carbonate Tunlonggonba Formation, lain over the volcanosedimentary Kwudi formation. The latter, at the top, has a limestone interbed, bearing the Yakhtashian-Bolorian fusulinids (Liang *et al.*, 1983) In the Shaksgam River Basin of Karakorum, the Kubergandian *Parafusulina* occurs in the middle part of the Shaksgam Formation (Desio *et al.*, 1991). This form associates with *Monodiexodina*, which is similar to the form described in the transgressive Amb Formation of the Salt Range (Douglas, 1970). Therefore, this latter must also be of the Kubergandian-Murgabian age.

In southwestern Pamirs, the deposits of the Kubergandian stage have gradual transitions to the Bolorian strata. In contrast to the latter, they have flyschoid features, indirectly evidencing the transgression and the deepening of the basin.

As we already have seen, the Kubergandian deposits of the most complete Tesak section in southern Afghanistan are separated from the top of Sakmarian strata only by a thin limestone bed, bearing Bolorian fusulinids. In the Vardak section, sandstones at the base of post-Sakmarian transgressive strata probably are also of the Kubergandian age. The Kubergandian-Murgabian fusulinids from that series basal strata were recognized by A. Siehl (1969) in the Tangi-Garu, Kamar, and Altimur sections.

A similar situation can be observed in sections of central Iran. In some of them, the Yakhtashian and Bolorian deposits are absent, and the transgressive carbonate strata begin with Kubergandian beds. One can draw this conclusion, although tentatively, using the published data (Taraz 1974; Ruttner *et al.*, 1968; Iranian-Japanese Research Group, 1981). More exactly, it is evidenced by samples from the Darai-Darchali section (the Abadeh district), which I have studied courtesy of Dr. D. Bakhbani. At the very base of the transgressive Surmag Formation, right above the Asselian-Sakmarian Vazhnan strata, there are abundant *Parafusulina* and *Skinnerella* typical of the lower Kubergandian. In passing somewhat higher, one can see the upper Kubergandian *Cancellina*, and then the *Praesumatrina* and *Neoschwagerina* of the lower Murgabian stage.

In all Permian sections of Transcaucasia, except the Saripap section mentioned above, the transgressive carbonate strata begin with Kubergandian, or sometimes Murgabian beds, which overlie Devonian and Lower Carboniferous deposits either with, or without, basal bauxites and red sandstones (Leven, 1975). The strata are well dated by their abundant fusulinids, such as: *Misellina ovalis*, *Armenina karinae*, *Cancellina*, and *Presumatrina*.

The Kubergandian deposits are unknown in the more western Tethyan regions. Murgabian deposits have been described in many Mediterranian regions. However, a thorough analysis of fusulinids, used in the age determination indicates that we are dealing with the strata of the Midian stage in its original definition (Leven, 1980), i.e., when the Arpin Formation of the Armenian stratotype is included in it. My recent revision of fusulinid assemblages at this stratigraphic level (Leven, 1992) has shown that, in this very place, appear characteristic Midian *Yabeina*, *Metadoliolina*, *Kahlerina*, *Codonofusiella*, *Dunbarula*, *Reichelina* as well as some typically Midian smaller foraminiferas: *Abadehella*, *Dagmarita*, *Neoendothyra*, *Paraglobivalvulina*, *Sphaerionia*, and many other forms. In association with them, there are also *Neoschwagerina margaritae*, *N. haydeni*, *Afghanella robbinsae*, and *Sumatrina annae*, the species used to classify their strata as *N. margaritae* zone of the Murgabian Stage. Thus, chronologically, this zone corresponds to the strata bearing first *Yabeina* forms and other Midian fusulinids, which creates a problem for its biostratigraphic identification. I am in favor of using the Midian stage, paying attention to the new form appearance in the foraminifer community. An important additional argument is the strata position of the beds at the base of the thick carbonate strata resulting from the wide Midian transgression, and spreading far beyond the distribution area of the Kubergandian-Murgabian deposits.

MIDIAN STAGE

The Midian transgression expanded all over the Tethys, and dominantly carbonate deposits of this age became most widespread. They overlie either Murgabian strata, or all stratigraphic units, including Precambrian ones.

We shall start the survey of the Midian Stage from Japan. The most important one here is the section of the Akasaka vicinity, where the biostratigraphic zonation of the Murgabian stage was suggested for the first time, using, among others, the *N. margaritae* Zone. The revision of all the data available for the section led me to the following conclusions. First, in the *N. margaritae* Zone, the proper index-species are absent, occurring only higher in the *Yabeina globosa* Zone of the section. Therefore, this species cannot be used as the zonal form, Second, *Yabeina* forms make their first appearance directly in this zone strata: *Yabeina ozawai*, and the form described by E. Ozawa (1927) as *N. margaritae*. The first *Metadoliolina* forms also appear here.

This is why S. Honjo had to rename the *N. margaritae* Zone as the *Y. ozawai* zone.

The *Y. ozawai* Zone, taken now for further discussion as the Midian one, demonstrates a gradual transition to underlying Murgabian deposits. In the Kitakami Mountains, the relationships between the Midian Stage and older strata are essentially different. The stage is represented here by a thick terrigenous formation whose basal conglomerates overlie eroded Bolorian limestones. The lower part of the formation is defined by Japanese researchers (Choi, 1973) as the *Monodiexodina matsubaishi* zone comparable with Kubergandian and Murgabian stages, yet it is more correct to regard it as the Midian strata. This is supported by the appearance of *Codonofusiella* and *Rauserella* forms in it, as well as by the similarity of the zone species with Midian *Monodiexodina* forms from Primor'e. Both Kubergandian and Murgabian stages seem to be missing in this section.

In Primor'e, the dominantly carbonate marine deposits of the Chandalasian Horizon correspond to the Midian Stage. Since the underlying sediments of the Vladivostok Horizon are mainly of a continental type, one can suggest a transgression during the Midian.

In southern China, the tightly related Midian and Murgabian deposits form the Maokou Formation. To the north, in the Chinling Range, there probably is a hiatus between the Midian and underlying deposits. The latter are dated here as the base of the Murgabian stage according to the occurrence of *Presumatrina* and *Cancellina* forms. Above them, there appear *Neoschwagerina margaritae*, *Sumatrina longissima*, and *Rauserella minima*, which are typical of the base of the Midian Stage according to my definition. In this case, the main part of the Murgabian Stage is missing here (Wang, Sun, 1973). The Midian deposits can be traced to the west from Chinling to eastern Kunlun Shan, and then probably to Altyndag and southern Kunlun Shan. I have no data on these sections and their relationships to underlying strata.

In the Karakul' zone of northern Pamirs, the Midian Stage is poorly studied, though there are fusulinids of this age detected in the upper limestones of the Balyandkiik Formation (Leven, 1967). In the Darvaz, the Midian deposits belong to the terrigenous Kaftarmol and carbonate Kafirbacha formations, transgressively overlying the red terrestrial sandstones and conglomerates of the Murgabian age. Their Midian age is proved by the conodont occurrence (*Stepanovites inflatus*) in the limestones of the Kafirbacha Formation (Leven, Kotsur, 1977).

In northern Afghanistan, the Midian stage is present in the Bamian and Bulola sections (Leven, 1982). Deposits of this age have basal conglomerates, lying over the eroded fusulinid limestones of the lower Murgabian age. Upsection, the conglomerates grade into limestones and shales, which turn quickly to clayey and sandy limestones with massive reef limestones at the top. In the lower part of the latter, and in clayey limestones beneath, there are fusulinids: *Kahlerina*, *Dunbarula*, *Codonofusiella*, *Neoschwagerina haydeni*, *Afghanella robbinsae*, *Sumatrina annae*, and characteristic small foraminifera of the *Abadehella* genus, all of them being typical of the Midian Stage base.

The Midian transgression is well pronounced in peri-Gondwanan sections of eastern Tethys. In southern Tibet, the Midian limestones of the Longge Formation with *Kahlerina* and *Yabeina* forms have erosional contact with Kubergandian Tunloggonba formation (Liang, *et al.*, 1983). In the Chapursan River valley (Karakorum), the member "3" of the Pandshah Formation rests on Sakmarian *Pseudofusulina*-bearing (Caetani *et al.*, 1990) limestones, and contain *Rectostipulina quadrata*, a peculiar foraminifer species so far unknown in deposits older than Midian. The Upper Midian *Gondolella bitteri* occurs in the higher strata of this section. Given all these data, one may suggest Midian age of member "3", and an essential hiatus between it and Sakmarian strata underneath.

In Kashmir, the lower limestones of the Zewan Formation belong to the Midian Stage and transgressively overlap siliceous and carbonaceous shales of shallow-water lagoon origin (the Ganga-mopteris beds) (Nakazawa *et al.*, 1975). The Zewan Limestone correlates with the Kuling Formation of the northwestern Himalaya region, which differs sharply with older deposits, including those of pre-Carboniferous age (Gaetani *et al.*, 1990 B). The Vargal Formation of the Salt Range is also of the Midian Stage, and bears *Reichelina* and *Noeschwagerina cf. margaritae* at the base. There is a discordance between it and the underlying Amb Formation, where the Murgabian Stage is partly, or entirely missing (Nakazawa, Dickins, 1986).

The lower Midian fusulinids occur in calcareous breccias-conglomerates of the Karas Formation of southeastern Pamirs, which has a sharp, probably erosional contact with underlying cherts. The range of the suggested hiatus is unknown, but seems to be limited (Leven, 1967). In contrast, the hiatus in sedimentation within the Bartang-Rangkul zone of central Pamirs is equal to the Yakhtashian-Murgabian time range, including both stages. As established recently, the limestones and dolomites of the Kyrk-Kochu Formation overlying the Sakmarian fusulinid limestones of the Dangikalon Formation, contain a foraminifer complex typical of the lower Midian strata, which is indicated by *Dunbarula*, *Codonofusiella*, *Yangchienia*, *Kahlerina*, *Chusenella*, *Dagmarita*, *Abadehella*, *Sphaerionia*, and many other forms present in their lower beds. The rocks are separated from the Sakmarian strata by a horizon of clastic rocks and bauxites.

In the central parts of Helmand and Argandab depressions of southern Afghanistan (Karapetov, Leven, 1973), the Midian deposits are undivided from the Murgabian strata, together forming the thick carbonate Chokhan group. In passing to depression margins, the Lower Permian and, then, Kubergandian-Murgabian strata gradually thin out, and Permian sections of interbasinal rises begin occasionally, once with

Midian deposits. In such cases, they overlie various older strata including those of the Precambrian.

In Transcaucasia, as well as in well studied sections of Elburz Mountains and central Iran, the Midian Stage is again undivided from the Murgabian one, being a part of single carbonate strata. A thin conglomerate layer is described only in the Djamal Formation near the Tabas, and seems to be a sign of short gap in sedimentation (Jenny-Deshusses, 1983). It is also possible that, in the Rutbarak and Rute sections of Elburz Mountains, the carbonate Rute formation begins not with the Murgabian stage, as it is assumed, but with the Midian strata. This may be evidenced by the greatly reduced thickness of formation at the first locality, and by the *Codonofusiella* occurrence in basal horizons of the second section (Jenny-Deshusses, 1983).

The Midian limestones overlie crystalline Precambrian rocks in Oman. Their age is defined by algae characteristic of *Permocalcus* genera, and by small foraminifers *Langella ocarina* occurring in the lower parts of the section. Higher up, there are typical Midian *Codonofusiella*, *Dunbarula*, and *Hemigordiopsis renzi* (Lys, 1986).

The Midian transgression is perfectly manifested in the more western Tethyan regions. In the eastern Taurus, near Tufanbeyli, Midian foraminifers *Dagmarita chanakchensis* were found in the lower beds of carbonate strata, covering, by their basal bauxites, Carboniferous limestones. Somewhat upsection, they contain *Shantia amosi*, *Paraglobivalvulina mira*, and *Stipulina* of the Midian-Dzhulfian age (Zaninetti *et al.*, 1981). In the Bademli section near Alanya, Midian *Basalina pulchra*, *Codonofusiella nana*, and *Sumatrina annae* were noticed at the base of limestone sequence above Asselian-Sakmarian deposits (Lys, 1986). In the Pamuchak section near Antalya, above the Otean continental sandstones, there are carbonate strata containing, at the base, diverse forms of *Codonofusiella*, and typically Midian *Hemigordiopsis renzi* (Lys, Marcoux, 1978). In eastern Anatolia (the Hazro section), the Midian *Chusenella cheni*, and *Paradunbarula matei* were found at the base of limestones above continental sandstones of tentatively Lower Permian age (Fontaine *et al.*, 1980).

The Midian limestones *in situ*, or as clasts in younger formations, are known to be on Cyprus, on some islands of the Aegean Sea, in the Apennines, and in Sicily. Their relations to older rocks remain unclear. However, the fact that Kubergandian and Murgabian sediments have not been found there may imply a gap in sedimentation of those times, or the originally limited distribution of corresponding deposits.

The Midian limestones and dolomites are known in several localities along the northeastern coast of the Adriatic Sea. The most complete data on them have been obtained in the Velebit and Lika ranges (Kochansky-Devidé, 1965). The thick dolomite strata with limestone interbeds overlie the clastic rocks (the Koshna Conglomerates) of the supposedly Early Permian age. Clearly Midian fusulinids, including *Yabeina syrtalis*, occur in the "second limestone zone," approximately in their middle part. In my opinion, the underlying "light-colored crystalline dolomites" are also of Midian age, as indicated by the occurrence of *Dunbarula*, *Kahlerina*, and advanced *Neoschwagerina* forms, one of which was classified as *N. margaritae*, but also could be a form of the Yabeina genus. In the "dark dolomites" from the lower part of the section, there are only Staffelidae forms, including *Sphaerulina croatica*, which occurs in the lower Midian strata of Afghanistan and Pamir. Finally, in the "first limestone zone", at the very base of dolomite strata, the *Chusenella* forms are in assemblage with *Reichelina* and *Colaniella nana*. Though, they have been defined tentatively (Kochansky-Davidé, personal communication), the Midian age of the forms is proved by a parallel occurrence of *Permocalcus* algae.

Northeastward, northward, and northwestward from the Adriatic coast, the marine carbonate facies of the Midian Stage are replaced by coastal continental and red continental sandstones (the Groeden, and Val Gardena formations). Similarly to carbonate Midian sediments, these deposits overlie transgressively and discordantly lower Permian formations (Massari, 1986). Their lower boundary seems to be diachronous, becoming younger when approaching basin margins.

The westernmost outcrops of marine Permian deposits of the Tethys are situated in the Jebel Tebaga Elevation of southern Tunisia. The lower part of the section consists of red shales and sandstones. Upsection, they grade into lightly colored quartzites and sandstones in which limestone, marl, and dolomite interbeds then appear. Playing an increasingly important role towards the top, the latter finally replace all terrigenous rocks. This transgressive strata is as thick as many hundreds of meters. Its lowermost limestones already contain Midian *Kahlerina* and advanced *Neoschwagerina* forms. Above, *Yabeina* forms appear to be associated with large *Dunbarula mathieui* (Skinner, Wilde, 1967).

DZHULFIAN AND DORASHAMIAN STAGES

As we have seen, the wide Midian transgression resulted in similar environments almost throughout the Tethys, and left behind widespread limestones with easily recognizable fusulinid assemblages of the same type. The situation began to change before the beginning of the Dzhulfian age. In many regions, the clayey carbonate sediments with cephalopod-brachiopod biofacies were replacing previous carbonate formations characterized by algae-foraminiferae biotas. It was the time of extinction for most fusulinids, happening even in regions where facial environments resembled those of the Midian age, as indicated by the continued deposition of algal limestones. The reasons of that extinction remain unclear. Moreover, it also affected some other fossil groups, e.g., cephalopod and corals, but any discussion on this problem cannot be considered more

than limited speculation. The mere fact, however, means that the Midian Dzhulfian boundary marks the beginning of the "great Late Permian extinction" continued in Dzhulfian and Dorashamian times, and ended by the beginning of the Triassic Period. Accordingly, one must seek a solution of this biotic problem within that time range, but not at the Permian Triassic boundary, which usually attracts the attention of scientists.

There are three reasons limiting the number of sections where one can detect boundaries between Midian, Dzhulfian, and Dorashamian stages: (1) the extinction of higher fusulinids guided the biostratigraphic correlation; (2) the rare occurrence of ammonoids and conodonts; (3) the more pronounced facies diversity than before. Because of this, it is difficult to understand general evolutionary trends of the Tethys basin, and to outline transgressive-regressive cycles that happened all over the basin.

Before starting the review of Dzhulfian-Dorashamian deposits, we need to clarify the disputable definition of the boundaries of two uppermost Permian stages. I prefer the scale adopted in the former Soviet Union (Leven, 1980). According to it, the Dzhulfian stage begins with the cephalopod zone of *Araxoceras latum*, which corresponds approximately to the base of the *Gondolella leveni* conodont zone. In this case, the beds with *Codonofusiella* and *Lepidolina kumaensis* still belong to the Midian stage. The lower boundary of the Dorashamian stage coincides with the base of the Phisonites *triangulus* zone. The beginning of the Triassic System is defined by the appearance of *Hindeodus parvus*.

The Upper Permian sections in Japan are poorly represented, thus giving little for solving the problems under consideration. The known Dzhulfian-Dorashamian deposits belong here to thick terrigenous formations, gradually replacing the upper Midian and terrigenous sediments with *L. kumaensis* (the Kuma, and Kitakami sections) (Kanmera, 1954; Choi, 1973). In the Maizuru section, there are limestone interbeds in the Dorashamian Stage (Ishii *et al.*, 1975). The Triassic deposits occur mainly as overlapping the erosion surface. Taking into account the quick facies changes along the strike, one can hardly suggest anything definite to outline general evolutionary trends in sedimentation. In the summary volume on Japanese geology (Minato *et al.*, 1965), the authors point to the Late Permian uplifts of Japan, especially within areas of the present-day Sea of Japan. They are evidenced by thick conglomerate formations abundant in sections since the end of Midian Age (the beds with *Lepidolina kumaensis*).

The late Midian-Dzhulfian uplifts occurred also in southern China, especially in its southeastern and western regions (Sheng *et al.*, 1985; Zhou, 1986) where coal seams were deposited. In the west, the uplifts were accompanied by basalt eruptions. The Dorashamian (Changxingian) deposits show transgressive overlapping, and consist of carbonate, carbonate-cherity, and clayey facies. At the very end of the Permian, the transgression shifted to a quick regression, and the Triassic formations were deposited almost everywhere in southern China above the stratigraphic unconformity.

The Dorashamian transgression was of limited scale, and did not touch the northwestern regions of China, where the marine sedimentation ceased at the end of the Midian Stage.

The late Midian uplifts were manifested within the northern Pamirs as well. In its Karakul' zone, the Murgabian-Midian limestones of the Balyandkiik Formation were replaced by clastic rocks of the Baigashka Formation. More to the west, in Darvaz, the Midian limestones of the Kafirbacha Formation lower subformation are overlain by variegared clays and marls of its upper member, this can also be considered as a result of uprising movements. At the very end of the Permian, they resulted in the regional desiccation, as evidenced by the unconformity at the base of Triassic deposits (Leven, 1967; Leven, Shcherbovich, 1978).

In northern Afghanistan, sediments of the age in question are unknown. It is likely, that there was no sedimentation due to the late Midian uplifts.

In the peri-Gondwanan part of the eastern Tethys, the first half of the time range under discussion was dominated by normal marine sedimentation. Traces of the Late Midian regression were not found anywhere. As an exception, the Salt Range section can be mentioned, where the Vargal Limestone of the Midian age are replaced, towards the top, by Dzhulfian sandstones (the Chkhidru Fm.) (Nakazawa, Dickins, 1986). Beyond this case, the Dzhulfian deposits are tightly related to Midian sediments, being represented by carbonate (southern Afghanistan), carbonate-cherty (southern Pamirs, Hunza River valley of Karakorum), and carbonate-terrigenous (Kashmir, Zanskar) facies. The definitely Dorashamian deposits proved to be present only in southeastern Pamirs (Davydov *et al.*, 1991), where they were recognized in the middle and upper parts of Takhtabulak Formation, consisting of tufaceous and clayey-marly rocks. Analogously, the uppermost clays and marls of the Kindil Formation in the Chapursan River valley (Karakorum) (Gaetani *et al.*, 1990 A) are assumed to be of the same age. It is possible that Dorashamian deposits, so far unrecognized, are present in a number of sections of southern Afghanistan. One also cannot dismiss the fact that they are missing in most of the sections as a consequence of the late Permian regression in this area. However, they did not reach the Himalayas, where in several localities (Kashmir, Spiti, Zanskar) the Permian strata are gradually progressing to Triassic deposits.

The upper Permian units are present in most of the west Tethyan sections. Perfect examples are observed in Transcaucasia, Iranian Azerbaijan, and central Iran (Abadeh). The deposits occur as condensed clay-carbonate sequences of the shallow-water quiet-lagoon sedimentation (Iranian-Japanese Research Group, The Permian and the Lower Triassic Systems in the Abadeh Region, Central Iran, 1981). They have a gentle transi-

tion to underlying sediments. There is no hiatus in passing to Triassic strata as well, but it can be cryptic. A similar situation seems to occur in the Zagros sections, where Dorashamian *Palaeofusulina*-bearing limestones are overlain, without visible unconformity, by Triassic strata with *Claraia*. A more abrupt transition to Triassic sediments is noted in the Elburz and eastern Iran (Tabas), where Dorashamian, and occasionally Dzhulfian stages can be missing (Rutner *et al.*, 1968; Stepanov *et al.*, 1969). In general, the Iranian and Transcaucasian sections evidence regression environments during sedimentation. The regression slowly evolved, and ceased at the boundary between the Permian and Triassic.

In sections of Oman and Turkey (Taurus, Antalya), the Midian limestones grade into the same facies of Dorashamian and Dzhulfian stages and, together, form a united sequence. The base of Triassic strata also consists of carbonate rocks (dolomites or oolitic limestones) which overlie sediments below without visible discordance. Features indicating a hiatus comparable to Dorashamian time range, or to its upper part, can be observed in the Pamuchak section near Antalya (Marcoux, Baud, 1986). Generally, gradual transitions from Midian to Dzhulfian deposits are disturbed in the eastern Anatolian Hazro section. Above the Midian limestones (the A-member of the Gomaniibrik Formation lower part), there are terrestrial sandstones and siltstones with coal interlayers (the B-member). Their age corresponds to the Dzhulfian, according to the position in the section. In passing higher, there is a C-member of Dorashamian limestones, and then the uppermost siltstones of the D-member, which are overlain by Triassic clayey deposits (Lys, 1986). It is easy to note that this transgressive-regressive succession coincides with that in southern China.

In the more western Tethyan regions, the deposits of the age under consideration occur in the Velebit and Lika sections. Similar to Midian sediments below, they are represented by dolomites and, to a lesser extent, by limestones, being overlain unconformably by Triassic dolomites (Kochansky-Devidé, 1965). The rocks have poor faunal characteristics, and the duration of the pre-Triassic hiatus remains unclear.

A different situation is observed within regions, situated in the Permian time along southern boundary of the European Hercynides (the Dolomitic and Carnic Alps, Karawanken, West Serbia, and northeastern Hungary). Like the Midian, the Dzhulfian deposits are represented partly, or entirely, by terrigenous red formations (the Groeden and Val Gardena sandstones). Above them, there are dolomites intercalated with limestone, gypsum, and anhydrite beds (the Zazar and Bellerofont formations). The Dorashamian age of the deposits was defined by the occurrence of *Palaeofusulina nana* at their base in a section near to Ljubljana (Ramovš, Kochansky-Devidé, 1979). It is possible that the lower part of the transgressive carbonate formation is still somewhere of Dzhulfian age. It is worth noting that similar facies at the top of the Permian deposits can be observed in the most eastern part of southern Europe, namely in the northern Caucasus. The red conglomerates and sandstones of the Kutanga Formation, overlying the continental lower Permian deposits, may be correlated with the Groeden Sandstones of western Europe. The carbonate Nikitin and Urushten Formations above them bear *Palaeofusulina* forms and brachiopods, both corresponding to the fossil types in the Zazar Formation (Kotlyar *et al.*, 1983). As one can see, either here, or in the south of western Europe, the Dorashamian deposits are lying transgressively. The pre-Triassic, late Dorashamian regression, traceable almost everywhere in Tethyan sections, was not so intense in the regions under consideration. It is more reasonable to suggest very low sedimentation rates for this time, which increase again to follow a new and wide Triassic transgression.

CONCLUSION

The data and considerations presented above permit us to draw the following conclusions.

1. Two essential transgressive-regressive cycles, the Asselian-Sakmarian, which probably began in late Carboniferous, and the Yakhtashian-Dorashamian, are traceable in the Permian history of the Tethys basin (Fig. 1). The transgressions were irregular, stepwise progressing, and interrupted by short regression phases. One of the most expanded transgressions took place in the Sakmarian, and resulted in the wider occurrence of Sakmarian deposits as compared to Asselian sediments. These deposits often overlap the pre-Asselian Formations. The second transgressive-regressive cycle begins with the short Yakhtashian-Bolorian transgression after the extensive sea regression at the end of the Sakmarian. The Yakhtashian-Bolorian transgression spread over the most depressed land areas. This is why the corresponding deposits are locally distributed and, beyond the carbonate facies, have quite abundant terrigenous sediments in their sequences.

A quick transgressive expansion happened at the beginning of the Kubergandian, and then again in the Midian time. It is possible that each of these transgressive waves were preceded by a desiccation of some basinal areas, which can be traced here and there at the boundaries between Bolorian and Kubergandian, and then between the Murgabian and Midian stages. However, these regressions were short and local. The Late Midian-Dzhulfian regression and the Dorashamian transgression were also irregular, but more clearly pronounced. An event of general significance was the regression at the boundary between the Permian and Triassic periods.

2. There is a direct and clear connection between transgressions and regressions on the one hand, and the main turning points of the fusulinid evolution on the other. Therefore, the extinction of many previously flourished genera and species is related to the post-Sakmarian desiccation. The Yakhtashian-Bolorian fusulinid assemblages, related in time to the beginning of a

main turning points of the fusulinid evolution on the other. Therefore, the extinction of many previously flourished genera and species is related to the post-Sakmarian desiccation. The Yakhtashian-Bolorian fusulinid assemblages, related in time to the beginning of a new transgressive-regressive cycle, are significantly different from those of the Asselian-Sakmarian age, though succeeding them at the genus level. The main peculiarity of the assemblages is illustrated by the first forms of genera, families, and even fusulinid order, which later became dominant. First of all, these are *Pamirina* and *Misellina* forms, giving rise to the Neoschwagerinida order. There are also roots of other widely occurring Upper Permian genera, such as *Neofusulinella*, *Yangchienia*, *Parafusulina*, and *Eopolydiexodina*.

Such features as the rapid differentiation and flourishing in the Kubergandian of the Verbeekinida order, the growing populations, specific diversity, and expanding residence area of other large assemblages of generic and higher ranks, borne in the Yakhtashian and Bolorian stages, can be explained by the quickly expanding transgression. The next widest Midian transgression affected the fusulinid community less than the Kubergandian one. This is understandable because there were no desiccated areas large enough before, as at the end of the Sakmarian time. The previous time it was simply an expansion of the earlier existing Murgabian basin. Respectively, most large Murgabian fusulinid taxons pass into the Midian stage, where they are represented by the most advanced species. The renovation of the Midian fusulinid community was more distinct in the Schubertellida and Ozawainella orders. They have several new genera, including the aberrant, such as *Lantschichites*, *Codonofusiella*, *Reichelina*, and others. The Kahlerinidae family came into existence and became abundant.

As a most simple explanation for the mass fusulinid extinction at the end of the Midian, one could suggest a regression at that time. However, this seems to be insufficient because it happened within limited areas. Most likely, there were some other reasons for it, but as yet, are still unknown. The Dzhulfian fusulinid complex is poor and monotonous. Three new genera of the Dorashamian Stage, *Palaeofusulina*, *Parananlingella*, and *Gallowaiinella*, are the most large and advanced forms of the Schubertellida order, and their appearance is evidently related to a short transgression which occurred in southern China and southern Europe at the very end of the Permian. The wide, all-encompasing Tethyan regression at the boundary between the Permian and Triassic periods resulted in the total disappearance of fusulinids.

3. Transgressions and regressions were occurring mainly over the entire Tethyan basin, thus providing the temptation to explain these events by the eustatic sea-level fluctuations. However, the transgression curve of this study coincides neither with Veil's curve, nor with a more detailed graph plotted according to American materials (Ross, Ross, 1987). Beyond a doubt, the Tethyan transgressions and regressions were, in some cases, related to tectonic events, and most of all to the phases of Hercynian orogeny, which were not synchronous everywhere. It is likely that the observed features are the result of a superposition of the eustatic, and less global tectonic processes. Their relative significance can be defined only after a more precise global biostratigraphy and event correlation.

REFERENCES

A Guide to Field Excursions in Tangshan, Nanjing Inst. of Geol. and Paleontol. Acad. Sinica, Nanjing, 1979.

Asseretto, R., The Paleozoic Formations in Central Elbours, Iran, *Riv. Ital. Paleontol.*, vol. 69, no. 4, pp. 503 - 543.

Bensh, F.R., *Stratigrafiya i Fusulinidy Verkhnego Paleozoya Yuzhnoi Fergany* (Stratigraphy and Fusulinids of the Upper Paleozoic of Southern Fergana), Fan, Tashkent, 1972.

Chen, S., Fusulinidae of South China, Part 1, *Paleontol. Sinica, Ser. B*, 1934, vol. 4, fasc. 2.

Choi, D.R., Permian Fusulinids from the Setamai-Yahagi District, Southern Kitakami Mountains, N.E. Japan, *Fac. Sci. Hokkaido Univ., Ser. 4, Geol. and Mineral.*, vol. 16, no. 1, p. 133.

Davydov, V.I., Kotlyar, G.V., Ivanova, Zh.B., *et al.*, Analogs of the Dzhaulfian and Dorashamian Stages in Southeastern Pamirs, *Permskaya Sistema Zemnogo Shara*, Sverdlovsk, 1991, pp. 34 - 35.

Desio, A., Caporali, A., Rampini, L., *et al.*, Geodetic Geophysics, and Geology of the Upper Shaksgam Valley (North-East Karakorum), and South Sinkiang. Italian Expedition to the Karakorum 1988, *Sci. Rep.*, Milano, 1991, pp. 99 - 188.

Douglas, R.C., Morphologic Studies of Fusulinids from the Lower Permian of West Pakistan, *Geol. Surv. Prof., Paper 643-G*, 1970, pp. 1 - 111.

Dronov, V.I. and Leven, E.Ya., New Data on the Permian Deposits of Central Pamirs, *Izvestiya Vuzov, Geologiya i Razvedka*, 1971, no. 3, pp. 10 - 15.

Fontaine, J.M., Brunton, C., Lys, M., *et al.*, Données Nouvelles sur la Stratigraphie des Formations Paleozoiques de la Platforme Arabe dans Region d'Hazro (Turquie), *C. R. Acad. Sc., Ser. D.*, Paris, 1980, vol. 291, pp. 917 - 920.

Gaetani, M., Garzanti, E., Jadoul, F., *et al.*, The North Karakorum Side of the Central Asia Geopuzzle, *Bull. Geol. Soc. Am.*, 1990a, vol. 102, pp. 54 - 62.

Gaetani, M., Garzanti, E., and Tintori, A., Permo-Carboniferous Stratigraphy in SE Zanskar, and NW Lahul (NW Himalaya, India), *Ecolog. Geol. Helv.*, 1990b, vol. 83, no. 1, pp. 143 - 161.

Honjo, S., Neoschwagerinids from the Akasaka Limestone, *J. Fac. Sci., Hokkaido Univ.*, 1959, vol. 4, no. 10(1), pp. 111 - 161.

Iranian-Japanese Research Group, The Permian and the Lower Triassic Systems in the Abadeh Region, Central Iran, *Mem. Fac. Sci., Kyoto Univ., Geol. and Mineral.*, 1981, vol. 47, no. 2, pp. 61 - 133.

Ishii, K., Okimura, Y., and Nakazawa, K., On the Genus colaniella and Its Biostratigraphic Significance, *J. Geosciences, Osaka City Univ.*, 1975, vol. 19, art. 6, pp. 107 - 129.

Jenny-Deshusses, C., *Le Permien de l'Elbourz Central et Oriental (Iran). Stratigraphie et Micropaleontologie*, Thése No. 2103, Univ. de Geneve, Imprimarie POT, 1983.

Kahler, F., Iranische Fusulinidae, *Jahrb. Geol. B.-A.*, 1974, vol. 117, pp. 75 - 107.

Kahler, F., Die Fusuliniden der Dorud Formation im Djadjerud-Tal nördlich von Teheran (Iran), *Riv. Ital. Paleont.*, 1976, vol. 82, pp. 439 - 466.

Kahler, F. *Die Definition der Trogkofel-Stufe Unter-Perm*, Karnische Alpen, Sonderdruck aus Sonderheft 36, Carinthia II, 1980, pp. 255 - 258.

Kahler, F., *Fusuliniden aus Karbon und Perm der Karnischen Alpen und der Karawanken*, Sonderheft 41, Carinthia II, Klagenfurt, 1983.

Kanmera, K., Fusulinids from the Upper Permian Kuma Formation, Southern Kyushu, Japan, with Special Reference to the Fusulinid Zone in the Upper Permian of Japan, *Mem. Fac. Sci., Ser. D, Kyushu Univ.*, 1954, vol. 4, no. 1, pp. 1 - 38.

Kanmera, K., and Mikami, T., Succession and Sedimentary Features of the Lower Permian Sakamotozawa Formation, *Mem. Fac. Sci., Ser. D*, Kyushu Univ., vol. 14, no. 3, pp. 265 - 274.

Karapetov, S.S. and Leven, E.Ya., Upper Paleozoic Deposits of Central Afghanistan, *Byull. MOIP, Otd. Geol.*, 1973, vol. 48, no. 1, pp. 30 - 40.

Kochansky-Devidé, V., Die Fusuliniden aus dem Karbon und Perm im Velebit und in der Lika (Kroatien), *Unteres Perm. Palaeont. Jugoslavica*, 1959, Sv. 3, pp. 1 - 60.

Kochansky-Devidé, V., Die Fusuliniden Foraminiferen aus dem Karbon und Perm in Velebit und in der Lika (Kroatien), Mittleres und Oberes Perm, *Jugoslavenska Acad. Zn. Um. Acta Geol.*, 1965, vol. 5, pp. 101 - 148.

Kochansky-Devidé, V. and Milanovic, M., Unterpermische Fusuliniden und Kalklagen des Tara-Gebietes in der Mittleren Crna Gora (Montenegro), *Geoloski Vjestnik*, Zagreb, 1962, vol. 15, no. 1, pp. 155 - 227.

Kochansky-Devidé, V., *Jugoslav. Acad. Sci., personal communication.*

Kotlyar, G.V., Zakharov, Yu.D., Kochirkevich, B.V., *et al.*, *Pozdnepermskii Etap Evolyutsii Organicheskogo Mira. Dzhulfinskii i Dorashamskii Yarusy SSSR* (Late Permian Stage of the Organic Kingdom Evolution. Dzhulfian and Dorashamian Stages of the Soviet Union), Nauka, Leningrad, 1983.

Leven, E.Ya., Stratigraphy and Fusulinids of Permian Deposits in Pamirs, *Trudy GIN*, Nauka, Moscow, 1967, vol. 167.

Leven, E.Ya., Stratigraphy of Permian Deposits in Transcaucasia, *Sov. Geol.*, 1975, no. 1, pp. 96 - 100.

Leven, E.Ya., Bolorian Stage of Permian: Substantiation, Characteristics, Correlation, *Izv. AN SSSR, Ser. Geol.*, 1979, no. 1, pp. 53 - 65.

Leven, E.Ya., Yakhtashian Stage of Permian: Framework, Characteristics, Correlation, *Izv. AN SSSR, Ser. Geol.*, 1980, no. 8, pp. 50 - 60.

Leven, E.Ya., *Ob"yasnitel'naya Zapiska k Stratigraficheskoi Shkale Permskikh Otlozhenii Oblasti Tetis* (Explanatory Notes to the Stratigraphic Scale of Permian Deposits, Tethys Region), VSEGEI, Leningrad, 1980.

Leven, E.Ya., Upper Paleozoic of the Charymdara, Gundara, and Zidadara River Basins (Southwestern Darvaz), *Byul. MOIP*, 1981, vol. 56, no. 4, pp. 40 - 52.

Leven, E.Ya., Cancellina Hayden Genus and Its Position in the Higher Fusulinid System, *Vopr. Mikropaleontologii*, 1982, no. 25, pp. 40 - 50.

Leven, E.Ya., The Sumatrinida Phylogeny and Problems of Permian Murgabian and Midian Stages Subdivision into Zones, *Paleontologicheskii Zhurn.*, 1992, no. 3, pp. 13 - 20.

Leven, E.Ya. and Kotsur, Kh., First Occurrence of Upper Permian Conodonts in Darvaz, *Dokl. AN SSSR*, 1977, vol. 234, no. 1, pp. 145 - 147.

Leven, E.Ya. and Shcherbovich, S.F., *Fusulinidy i Stratigrafiya Assel'skogo Yarusa Darvaza* (Fusulinids and Asselian Stage Stratigraphy in Darvaz), Nauka, Moscow, 1978.

Leven, E.Ya., Les Gissements Permien et les Fusulinides de l'Afghanistan du Nord, *Notes et Memoires sur le Moyen-Orient*, 1971, vol. 12, part 1, pp. 1 - 35.

Liang, D., Nie, Z., Guo, T., *et al.*, Permo-Carboniferous Gondwana-Tethys Facies in Southern Karakoran Ali, Xizang (Tibet), *J. Wuhan Coll. Geol.*, 1983, no. 1, pp. 9 - 27.

Lys, M., *Biostratigraphie du Carbonifere et du Permien en Mesogee*, Thése. Univ. de Paris-Sud. Centre d'Orsay, 1986.

Lys, M., and Lapparent, F., Foraminiferes Microphacies du Permien de l'Afghanistan Central, *Notes et Memoires sur le Moyen-Orient*, 1971, vol. 12, part 1, pp. 49 - 133.

Lys, M., Marcoux, J., Les Niveaux du Permien Superieur des Nappes d'Antalya (Taurides Occidentales, Turquie), *C.R. Acad. Sci., Ser. D*, Paris, 1978, vol. 286, pp. 1417 - 1420.

Marcoux, J. and Baud, A., The Permo-Triassic Boundary in the Antalya Nappes (Western Taurides, Turkey), *Mem. Soc. Geol. It.*, 1986, vol. 34, pp. 243 - 252.

Massari, F., Some Thoughts on the Permo-Triassic Evolution of the South-Alpine Area (Italy), *Mem. Soc. Geol. It.*, 1986, vol. 34, pp. 179 - 188.

Minato, M., Gorai, M., and Hunahashi, M., (Eds.) *The Geologic Development of the Japanese Islands*, Tsukiju Shokan, 1965.

Nakazawa, K., Kapoor, H.M., Ishii, K., *et al.*, The Upper Permian and the Lower Triassic in Kashmir, India, Mem. Fac. Sci., Kyoto Univ., *Geol. Mineral.*, 1975, vol. 42, no. 1, pp. 1 - 106.

Nakazawa, K. and Dickins, J.M. (Eds.), *The Tethys*, Tokai Univ., Tokyo, 1986.

Nogami, I., Permische Fusuliniden aus dem Atetsu-Plateau Sud-West Japans, *Mem. Col. Sci., Ser. B*, Kyoto Univ., 1961, vol. 27, no. 3, pp. 159 - 248.

Ota, M., Geological Studies of Akiyoshi. Part 1. General Geology of the Akiyoshi Limestone Group, *Bull. Akiyoshi-dai Museum Nat. History*, 1977, no. 12, pp. 1 - 51.

Ozawa, T., Kobayashi, F., and Watanabe, R., Biostratigraphic Zonation of Late Carboniferous to Early Permian Sequence of the Akiyoshi Limestone Group, Japan, and Its Correlation with Reference Sections in the Tethyan Region, *Saito Ho-on Kai Spec. Pub.*, 1991, no. 3, pp. 327 - 341.

Ozawa, Y., Stratigraphical Studies of the Fusuline Limestone of Akasaka, Province of Mino, *J. Coll. Sci. Imperial Univ. Tokyo*, 1927, vol. 2, no. 3, pp. 121 - 164.

Ramovš, A., Razvoj Zgornjeca Perma v Loških in Polhograjskih Hribit, *Razpr. Slov. Acad. Znan. Imet.*, Ljubljana, 1958, vol. 4, no. 2. pp. 455 - 622.

Ramovš, A., and Kochansky-Devidé, V., Carboniferous and Permian Beds from the Northern Julian Alps, *Geologija*, 1979, vol. 22, no. 1, pp. 21 - 54.

Reichel, M., Unterpermische Fusuliniden aus dem Karakorum und dem Aghil-Gebirge, Visser C., Ed., *Wiss. Ergeb. Exped. Karakorum*, 1940, vol. 3, pp. 89 - 118.

Ross, Ch.A., and Ross, J.R.P., Late Paleozoic Sea Levels and Depositional Sequences, *Cuchman Found. Foram. Research*, 1987, Spec. Publ. 24, pp. 137 - 148.

Rui, L. and Hou, J.H., *Late Carboniferous Fusuliniceans from Southeastern Shanxi, Late Paleozoic Coal-Bearing Strata and Biota from Southeastern Shanxi*, China, Nanjing Univ., 1987, pp. 139 - 280.

Rui, L., Zhang, L., and Wang, Zh., Notes on the Carboniferous-Permian Boundary in South China, XI Congres Internat. de Stratigraphie et de Geologie du Carbonifere, Beijing, 1987, *Compt. Rend.*, 1989, no. 2, pp. 239 - 250.

Ruttner, A., Nabavi, M.H., and Hajian, J., Geology of the Shirgesht Area (Tabas Area, East Iran), *Geol. Surv. Iran*, 1968, Rep. 4.

Ruzhentsev, S.V., Pospelov, I.I., and Badarch, G., Tectonics of the Mongolian Indosinids, *Geotektonika*, 1989, no. 6, pp. 13 - 23.

Saito, Y., Geology of the Younger Paleozoic System of the Southern Kitakami Massif, Iwate Prefecture, Japan, *Sci. Rep. Tohoku Univ., Ser. 2*, 1968, vol. 40, no. 2, pp. 79 - 139.

Sheng, J., Rui, L., and Chen, Ch., *Permian and Triassic Sedimentary Facies and Paleogeography of South China, The Tethys*, Tokai Univ., 1985, pp. 59 - 80.

Siehl, A., Zur Stratigraphie und Palaogeographie des Perm in Afghanistan, *Geol. Rundschau*, 1969, vol. 56, pp. 795 - 812.

Silva, I.P., *Permian Foraminifera from the Upper Hunza Valley, Italian. Exp. Karakorum and Hindu Kush*, Sci. Rep., Leiden, 1965, pp. 90 - 125.

Skinner, J.W., and Wilde, G.L., Permian Foraminifera from Tunisia, *Univ. Kansas Paleontol. Contr.*, 1967, paper 30.

Stepanov, D.L., Golshani, F., and Stöcklin, J., Upper Permian and Permian-Triassic Boundary in North Iran, *Rep. Geol. Survey Iran*, 1969, no. 12.

Taraz, H., Geology of the Surmaq-Deh Bid Area, Abadeh Region Central Iran, *Rep. Geol. Survey Iran*, no. 47, 1974.

Vachard, D., Thetys et Gondwana an Paleozoique Superieur les Donnees Afghanes, *Docum. Trav. IGAL*, Paris, 1980, no. 2.

Wang, J.H., Boundary and Zonation of the Chihsia Formation in Nankin, *Acta Stratigraphica Sinica*, 1978, vol. 2, no. 1, pp. 67 - 73.

Wang, K.L., and Sun, X.F., Carboniferous and Permian Foraminifera of the Chinling Range, and Its Geologic Significance, *Acta Geol. Sinica.*, 1973, no. 2, pp. 137 - 178.

Xiao, W., Zhang, L., Wang, H., *et al.*, *Early Permian Stratigraphy and Faunas in Southern Guizhou*, The People Publ. House of Guizhou, 1986.

Zaninetti, L., Altiner, D., and Catal, E., Foraminiferes et Biostratigraphie dans le Permien Superieur du Taurus Oriental, Turquie, *Notes Lab. Pal. Univ. Geneve*, 1981, vol. 7, no. 1, pp. 1 - 37.

Zhang, L.X. (Ed.), Carboniferous Stratigraphy in China, Contr. 11th Intern. Congress, Carbon. Stratigr., *Geology*, China Sci. Press, Beijing, 1987.

Zhang, Z., The Permian System in South China, *Newslett. Stratigr.*, 1984, vol. 13, no. 3, pp. 156 - 174.

Zhang, L.X., Upper Carboniferous Fusulinids of the Kelping and Nearby Xingiang Areas, *Acta Paleontol. Sinica*, 1963, vol. 11, no. 1, 2, pp. 55 - 63.

Zhou, Z., Two Ecological Patterns of the Permian Ammonoids, *Sci. Sinica, Ser. D*, 1986, vol. 29, no. 1, pp. 96 - 112.

Stratigraphy and Geological Correlation, Vol. 1, No. 1, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 1, 1993, pp. 76 - 86.
Original Russian Edition Copyright © 1993 by Stratigrafiya. Geologicheskaya Korrelyatsiya, Ivanova, Chuvashov.
English Translation Copyright © 1993 by Interperiodica.

Bashkirian Stage of the Sim Brachysyncline (the Western Slope of the South Urals)

R. M. Ivanova and B. I. Chuvashov

Institute of Geology and Geochemistry, Ural Branch, Russian Academy of Sciences, Pochtovyi Per. 7, Ekaterinburg, 620219 Russia

Manuscript received December 4, 1991

Abstract – Sections of the Bashkirian stage in the Sim brachysyncline were examined between Yuryuzan'-Ai and Belaya rivers basins on the western slope of the South Urals. Foraminifers were used as the basis for a biostratigraphic subdivision, and the boundary between the Lower and Middle Carboniferous was substantiated by them. This boundary corresponds to the base of the *Homoceras* goniatite zone. The Bashkirian was divided into two substages and six horizons analogous to those of the stratotype on the Askyn River. The five lower horizons are composed of carbonate rocks and the upper one (the Asatau) is terrigenous.

The northern part of the South Urals western slope, lying between the latitudinal segments of the Ufa and Belaya rivers, is the stratotype area of the Bashkirian that was defined by S.V. Semichatova in 1934 by brachiopods as the beds with fauna. The Bashkirian deposits are distributed everywhere in this region. They are constituents of the units composing the West Ural folded zone, the basement of the Ural foredeep within the boundaries of the Yuryuzan'-Ai and Belaya basins and spread over the margin of the East European Platform. However, there are too few investigations related to their detailed subdivision, and these investigations were done on the basis of brachiopods 20 - 60 years ago. For a long time the Bashkirian did not have an appropriate stratotype and generally recognized stratigraphic range, until (Z.A. Sinitsyna and I.I. Sinitsyn, 1987) published data about the geological section along the Askyn River where foraminifers were used as the basis for biostratigraphic subdivision (Groves, 1988).

The Sim brachysyncline, whose geological sections will be discussed later, is composed of the upper Paleozoic deposits and represents the part of the western side of the Ural foredeep between the Yuryuzan'-Ai and Belaya basins that became to some extent isolated as the result of post-Bashkirian tectonic and erosional processes. The brachysyncline is situated in the stratotype area and in the same (West Ural) structural facies zone with the stratotype of the Bashkirian on the Askyn River. This makes it possible to trace and substantiate all the boundaries of defined units by comparing them with the stratotype, as well as to supplement and specify their microfaunal characteristics.

The stratotype is represented mainly by algal, slimy fine- detrital, pelitomorphic, and dolomized limestones (Sinitsyna and Sinitsyn, 1987). The rocks lack foraminifers, whereas the sections in the vicinity of the Sim Factory are more favorable for foraminifers facies, with diverse and abundant microfauna and algoflora. This especially concerns the lower part of the Bashkirian and defines more reliably the boundary between the Lower and Middle Carboniferous and the boundaries between the substages and horizons of the Bashkirian.

Within the confines of the studied area in the western Urals (between the Ufa and Belaya rivers), one can recognize, from east to west, several types of the Bashkirian sequences (Table 1).

(1) The Uraim, the easternmost type, represented by carbonates and in the upper part by sandy-clayey rocks and conglomerates and stretching as a narrow belt along the Urals to the Aya River.

(2) The Lakly, composed at the bottom of limestones with minor dolomites and of siliceous clayey - marly rocks of the Kolosleika Formation at the top (Chuvashov and Mizens, 1990). This can be traced from the Ufa River to the Sim River.

(3) The Askyn type of the sequence, completely carbonate, is distributed in the western parts of the Yuryuzan'-Ai and Belaya rivers basins and in adjacent areas of the platform.

The sequences of the Sim brachysyncline belong to the Lakly type and are situated near the western limit of this type distribution area. The characteristic feature of these sequences is gradual replacement of the clayey-marly Kolosleika Formation from east to west by limestones. Of six studied sections in the brachysyncline (Fig. 1) only one is well known in literature. This is the most eastern section near Yakhino (Yakh'ya) Village and it crops out in the cutting of the Chelyabinsk-Ufa railway and along the Berdyaush River (Petrenko and Kiparisova, 1937; Einor, 1958). O.L. Einor divided the Bashkirian there into two horizons: C_2^1 – I and C_2^1 – II. In the vicinity of the Sim Factory "the Bashkirian Beds with *Choristites bisulcatiformis*" were schematically described by E.A. Dmitriev and D.V. Nalivkin (1927)

Table 1. Types of sequences of the Bashkirian in the Ufa Amphitheater

Stratigraphic scale			Types of sequences: Askyn	Lakli	Uraim
Middle Carboniferous	Bashkirian, Upper Substage	Moskovian: Vereya	**Tashly Formation** Micritic limestone, marlstone, cherts, rare interbeds of detrital limestone	**Kolosleika Formation** Dark-grey and black claystone, marlstone, and cherts with thin (3 - 10 cm) interbeds of micritic and, more rarely, organogenic-detrital limestones	Polymictic sandstone and marlstone; micritic, sandy, oolitic, and detrital limestones; lenses and interbeds of cherty-calcareous conglomerate
		Asatau	Bedded limestone: organogenic-detrital, oolitic with thin biostromes and bioherms of briozoan-algal (of Donezella) composition and stromatolites, lenses of brachiopodal and coral limestones		Uraim Formation
		Tashasta			
	Lower Substage	Askynbash			Bedded detrital, oolitic, rarely sandy and marly limestones, claystone, and marlstone
		Akavas			
		Syuran			
		Bogdanovka			
Lower	Serpukhovian		Organogenic-detrital, frequently dolomitized limestone, dolomite; the upper part in the east is composed of massive stromatolitic-briozoan bioherms		

on the slopes of the Zhukova Shishka Mountain N.P. Malakhova and A.A. Malakhov (1961) established in the same locality the lower Bashkirian, consisting of two horizons: C_2^1 – I and C_2^1 – II. Only the papers mentioned above contain information on the stratigraphy of the Bashkirian in the Sim brachysyncline.

The investigation of the Serpukhovian Stage in its full range was not our purpose, therefore, we only give a brief description of the Serpukhovian-Bashkirian boundary in sections I, II, IV, V and VI (Fig. 2). In the two last sections dolomites and silicified limestones dominate near the boundary between the Lower and Middle Carboniferous. They contain only rare specimens of *Globivalvulina* sp., therefore, the boundary between the two epoches was defined provisionally.

THE SERPUKHOVIAN STAGE

The Staroutkinskii Horizon. The upper horizon of the Lower Carboniferous, first identified in the Sim brachysyncline, is represented most completely in sections I, II, and IV. It is well exposed and consists of limestones of grey and light-grey colors, granular, partially silicified or dolomitized, brecciated, thick- and medium-bedded, with lens-like interbeds of brachiopodal, crinoidal, and foraminiferal coquinas, organogenic sandstone, and pelitomorphic and coralline limestone varieties. The foraminiferal interbeds are composed principally of *Tolipammina*, the brachiopodal, of *Striatifera*, *Gigantoproductus*, *Athyris adepressiora* Ein.[1] and others, and the coral interbeds of Lithostrotianidae. Besides *Tolipammina*, the most typical foraminifers are as follows: *Rhectoendothyra*, rare *Endothyranopsis crassa* (Brady), *Litoutubella* cf. *magna* Raus., *Pojarkovella* sp., Pseudoendothyrae, abundant *Eostaffellae*, including *Eostaffella ikensis tenebrosa* Viss., *E. postpriokensis* Vdov., *E. pseudostruvei angusta* Kir., *E. mirifica* Brazhn., *E. ovoidea statula* Reitl., *Eostaffellina paraprotvae* (Raus.), and others, *Mediocris mediocris* (Viss.), *Plectostaffella jakhensis* Reitl., large *Climacamminae*, *Endotaxis brazhnikovae* (Bog. et Juf.), *Biseriella parva* (N. Tchern.), *B. scaphoidea* (Reitl.), *Asteroarchaediscus subbaschkiricus* (Reitl.), *Ast. parvus* (Raus.), *Archaediscus convexus* Grosd. et Leb., *Neoarchaediscus* sp., *Rugosoarchaediscus tumefactus* (R. Ivan.); diverse algae: *Malakhovella* sp., *Asphaltinella horovitzi* Mamet et Roux, *Claracrusta catenoides* (Hom.), *Nanopora fragilissima* Masl., *Clavaporella reinae* Racz, *Fasciella kizilia* R. Ivan., rare *Donezella lutugini* Masl., *Stacheoides polytrematoides*

[1] For the Yakhino section all microfaunal determinations are given according to Einor (1958).

(Brady), and *Cuneiphycus jonhsoni* Flüg. The horizon thickness varies from 36 to 51 m.

Two kilometers southward of Zhukova Shishka Mountain, on the right rocky bank of the Sim River (section IV), the Staroutkinsk Horizon is made up of light dolomite with interbeds of pelitomorphic and fine-detrital limestones, lenses of light cherts, and abundant *Striatifera striata* and productid banks.

THE BASHKIRIAN STAGE

Within the Sim brachysyncline, the Bashkirian deposits were found to consist of five horizons: the Bogdanovka, Syuransk, Akavas, and Askynbash of the Lower Bashkirian and the Tashasta of the Upper Bashkirian. The upper Asatan Horizon (Table 1) is probably a part of the clayey-siliceous Kolosleika Formation (Chuvashov and Mizens, 1990).

The Bogdanovka Horizon was first established on the western slope of the South Urals in the Zilair megasynclinorium within the *Homoceras* goniatite zone (Einor *et al.*, 1973). It corresponds to the *Plectostaffella bogdanovkensis* foraminiferal zone (Reitlinger, 1980). The Bogdanovka Horizon in the Sim brachysyncline is defined for the first time and only in the Yakhino section (Dmitriev and Nalivkin, 1927) where it is quite different from the underlying horizon deposits, consisting of dark-grey, massive, medium-bedded limestones, pelitomorphic, foraminiferal fine-detrital, and oolitic. The following fossils were found in these limestones: *Tolipammina*, *Pseudoglomospira*, *Haplophragmina*, *Endothyra*, numerous and diverse *Pseudoendothyra* such as *Ps. preobrajenskyi* (Dutk.), *Ps. probatus* (Durk.), and *Ps. spectata* (Durk.), small *Mediocris*, *Endostaffella*, *Eostaffella postmosquensis acutiformis* Kir., *E. pseudostruvei angusta* Kir., *E. pseudoovoidea* Reitl., *E. klautzanae* Grozd. et Leb., *Millerella* (*Eostaffella*) *lenticula* (Grozd. et Leb.), *M. variabilis* Raus., *Plectomillerella* sp., *Plectostaffella varvariensis* Brazhn. et Pot., *Pl. varvariensis lata* Brazhn. et Vdov., *Pl. schwetzovi asymmetrica* Brazhn. et Vdov., *Pl. jakhensis* Reitl., *Pl. mira* Raus., *Pl. bogdanovkensis* Reitl., *Globivalvulina moderata* Reitl., *Gl. glomerata* R. Ivan., *Asteroarchaediscus* and *Neoarchaediscus*; algae: *Rectangulina tortuosa* Antrop., *Praedonezella cespeformis* Kul., *Anthracoporellopsis machaevi* Masl., *Nanopora fragilissima* Masl., *N. anglica* Wood., *Asphaltinella horowitzi* Mamet et Roux, *Beresella* sp., *Donezella lutugini* Masl., *Dvinella distorta* Kul., *Paraepimastopora* aff. *kansasensis* (John.), and *Stacheoides*; brachiopods: *Schizophoria resupinata* Mart., *Productus pseudogruenewaldti* Ein., and *Athyris ambiqua* Sow. Striatifera and Choristites are absent. The thickness of the Bogdanovka Horizon is 17 m.

Amongst foraminifers, *Pseudoendothyra* and *Plectostaffella* clearly dominate, especially *Plectostaffella varvariensis*, *Pl. jakhensis*, and a zonal species *Pl. bogdanovkensis*.

The Bogdanovka Horizon in the rest of the sections was not recognized because of different reasons: either the boundary between the Lower and Middle Carboniferous is not exposed, or the section base is made up of a thick dolomite member. Nevertheless, in the last case the boundary can be identified quite reliably by the change of carbonate rock types.

The Syuran Horizon corresponds to the stan- dard *Eostaffella postmosquensis-E. pseudostruvei-Plectostaffella varvariensis* foraminiferal zone, but in the Urals it seems more advisable to assume *Semistaffella variabilis* as the index species for this zone (Kulagina, 1990) because it is most justified from the evolutionary point of view. The first three species appear as far back as the upper part of the Serpukhovian and are characteristic of the Syuran Horizon within the ranges of the two cephalopod zones, *Homoceras* and *Reticuloceras*, whereas *Semistaffella variabilis* marks the beginning of Syuran time, is the ancestral species for the Middle Carboniferous *Pseudostaffella* genus, and corresponds only to the *Reticuloceràs* genozone.

Sediments of the Syuran Horizon were found in all the studied sections except sections III (outcrop 1724) and V (outcrop 1729 near Koloslei Village) where the section base is composed of dolomite without definable microfauna (Fig. 2). The horizon is represented chiefly by brown and dark-grey, less frequently grey, fine-grained, medium- and thick-bedded limestones, silicified in some places, foraminiferal and pelitomorphic, with interbeds of algal (of Donezella), crinoidal, and brachiopodal (of Martinia) limestones. The assemblages of foraminifers, algoflora and brachiopods are rather diverse: numerous attached forms (*Tolipammina*, *Palaeonubecularis*, *Ammovertella*), *Planoendothyra aljutovica* Reitl., *Endothyra* ex gr. *bowmani* Brazhn. et Pot., *Bradyina* (5 species), *Pseudoendothyra tchernjaevae* (Grozd. et Leb.), *Eostaffella pseudostruvei* Raus. et Bel., *E. postmosquensis* Kir. with subspecies, *E.pseudoovoidea* Reitl., *E. kashirica* Raus., *E. kashirica rhomboides* Raus., *E. ovoidea* Brazhn. et Pot., *E. korobcheevi* Raus., *Plectostaffella varvariensis* (Brazhn. et Pot.) with subspecies, *Pl. mira* (Raus.), *Pl. mira obtusa* Reitl., *Pl. jakhensis* Reitl., *Pl. bogdanovkensis* Reitl., *Pl. cuboides* (Rum.), *Millerella uralica* Kir., *M. umbilicata* Kir., *M. paraumbilicata* Man., *Semistaffella variabilis* Reitl., *S. primitiva* Reitl., *Climacammina aljutovica* Reitl., *Biseriella minima* (Reitl.), *B. scaphoidea* (Reitl.), *Globivalvulina kamensis* Reitl., *Gl. moderata* Reitl., *Gl. procera* Post., *Asteroarchaediscus subbaschkiricus* (Reitl.), *Ast. pustulus* (Reitl.), *Neoarchaediscus postrugosus* (Reitl.), *N. gregorii* (Dain.), *N. incertus* (Grozd. et Leb), *Rugosoarchaediscus tumefactus* (R. Ivan)., *Eolasiodiscus grandis* R. Ivan; algae: *Donezella lutugini* Masl., *Claracrusta catenoides* Hom., *Anthracoporella girtyi* Mamet et Roux, *Cuneiphycus jonhsoni* Flüg., *Stacheia marginulinoides* Brady, *Stacheoides polytrematoides* (Brady), *Aoujgalia variabilis* Term. et Term.; brachiopods: *Martinia semiconvexa* Chao, *M. timanica* Tchern., *Productus concinnus* Sow., *Pr. corrugatus*

McCoy, *Choristites bisulcatiformis* Semich.; tetracorals, and chaetetids.

Eostaffella, *Plectostaffella*, and star-shaped *Asteroarchaediscus* quantitatively dominate amongst the foraminifers. *Plectostafella varvariensis baisultanica* Reitl., *Pl. varvariensis indecora* Reitl., and *Pl. mira* (Raus.) are found more frequently than other species of the genus. The algae are various, but either red or green algae predominate. *Choristites* and *Productus* are few and distributed irregularly, whereas *Martinia* form large accumulations. The thickness of the Syuransk Horizon is from 36.2 to 50.5 m.

In the Yakhino section (1), we divided the C_2^1 – I Horizon identified by Einor (1958), into two horizons: the Bogdanovka with a zonal species *Plectostaffela bogdanovkensis* and the Syuransk with a zonal species *Semistaffella variabilis*. *Striatifera* and *Gigantoproductus* are absent in the C_2^1 – I Horizon.

Akavas Horizon has been identified in the Sim brachysyncline in all the studied sections according to the appearance of the zonal species *Pseudostaffella antiqua*, but its full sequence can be found in 4 sections: I, II, III and VI (Fig. 2). The lower part of the horizon is represented predominantly by dark-grey to black with brown tint algal and algal-polydetrital limestones containing the interbeds of oolitic limestone. The upper part is made up of grey and brown-grey brecciated silicified limestones with lenses of cherts, interbeds of dark pelitomorphic, microgrumous, algal and crinoidal-brachiopodal limestones. The characteristic feature of the Akavas Horizon is a presence of several interbeds of oolitic limestone and almost black coquina, (thickness – 0.8 m), consisting of *Productus concinnus* Sow., *Marginifera borealis* Iva., *Schizophoria resupinata* Mart., *Atthyris ambiqua* Sow., and other brachiopods. Brachiopods *Productus neoinflatus* Lich., *Pr. uralicus* Tchern., *Choristites pseudobisulcatus* Frcks. et Rot., *Ch. transwersus* Semich., *Ch. uralicus asiatica* Frcks., *Ch. mansuji* Chao, bryosoans, small crinoid columnals, and pelecypods *Pterinopecten* sp. and *Grammysiopsis* cf. *donaica* Tchern., sometimes form lens-shaped accumulations, and are dispersed across the entire sequence. The foraminifers and algae are diverse and numerous. The main members of the assemblage are *Tolipammina*, large *Bradyina*, *Eostaffella*, *Plectostaffella*, *Semistaffella*, *Pseudostaffella*, and *Archaediscus*. The majority of them, except *Eostaffella ljudmilae* Raus., *E. acutissima* Kir., *E. mixta* Raus., *Plectostaffella mira* (Raus.), and all of *Pseudostaffella* which are a perfectly new element of the community, forming about half of the entire microfaunal assemblage (*Pseudostaffella antiqua* (Dutk.), *Ps. korobezkikh* Raus. et Saf., *Ps. ziganica* Sin., *Ps. sofronizkyi* Saf., *Ps. varsanofievae* Raus., *Ps. irinovkensis* Leont., *Ps. minor* Raus. Ps. *antiqua grandis* Schlyk., *Ps. antiqua posterior* Saf., *Ps. compressa* Raus., *Ps. paracompressa* extensa Saf., *Ps. proozawai* Kir., *Ps.* ex gr. *praegorskyi* Raus., and *Ps. turbulenta* Grozd. et Leb.) move upward from the Bashkirian section into overlying deposits. Characteristic is the appearance of *Schubertella obscura* Lee et Chen and *Ozawainella aurora* Grozd. et Leb. and the presence of numerous *Bradyina pseudonautiliformis* Reitl. and *Br. eonautiliformis* Reitl with thick coarse-porous walls, *Endotaxis*, and attached forms in fine-grained limestone varieties. *Globivalvulina and Biseriella*, as well as flattened *Neoarchaediscus*, especially, *N. gregori* (Dain.), *N. incertus* (Grozd. et Leb.), too, gravitate toward this assemblage. The first *Ozawainella rhomboformis* Man. and *Oz.* ex gr. *paratingi* Man. have been found in the upper part of the horizon. Green and red algae are approximately equal in their diversity, but only *Donezella* and *Claracrusta* are rock-forming genera.

In the area of Simskaya railway station, downstream from the mouth of the Kolosleika River (section VI), the Akavas Horizon is made up of earthy-grey and yellowish-grey dolomites with rare interbeds of foraminiferal, detrital, and pelitomorphic limestones. The interbeds are usually grey or dark-grey with a brown tint and contain small productids, *Athyris* sp., Choristites sp., rare tetracorals, and crinoids. The upper boundary of the Akavas Horizon is marked here by the replacement of thin-bedded (5 - 10 cm) dark-brown limestone by earthy-grey massive dolomite. The horizon thickness is 25 - 50 m.

The Askynbash Horizon corresponds to the *Pseudostaffella praegorskyi-Profusulinella staffellaeformis* foraminiferal zone and was identified in five investigated sections: I, II, III, V, and VI (Fig. 2). The criterion for dividing the Akavas and Askynbash horizons is the appearance in the sequence of the index species *Pseudostaffela praegorskyi*.

The limestones of the Askynbash Horizon crop out either as small gently dipping ledges (the Yakhino section) or in the form of steep rocky cliffs (Zhukova Shishka). They are represented predominantly by fine-grained, dark- and brown-grey, bedded limestones, with few interbeds of lighter dolomite. Occasionally, a thin-bedded, schistose, cherty-clayey rock with beds 1 - 2 cm thick is present there. Black cherts and lenses of brachiopodal limestone have been found more often. Limestones, chiefly algal and algal-foraminiferal, with interbeds of pelitomorphic, microgrumous, and bituminous varieties are very similar to the rocks of the same age from the stratotype section. Foraminifers and algae are diverse and abundant but the main background of the assemblage is represented by *Pseudostaffella*. These are *Pseudostaffella antiqua* (Dutk.), *Ps. antiqua posterior* Saf., *Ps. varsanofievae* Raus., *Ps. irinovkensis* Leont., *Ps. korobezkikh* Raus. et Saf., *Ps. compressa* Raus., *Ps. paracompressa* Saf., *Ps. paracompressa* extensa Saf., *Ps. praegorskyi* Raus., *Ps. proosawai* Kir. f. *minima*, and *Ps. nibelensis* Raus. Notably, the flattened, lens-shaped forms with pointed median area, and often with contiguous whorls predominate among *Eostaffela* (i.e., *E. mutabilis* Rays., *E. mutabilis rjasanensis* Raus., and *E. korobcheevi* Raus). *Asteroarchaediscus* and *Neoarchaediscus* are present as frequently as before. The presence of ancient recurrent forms, such as *Bisphaera*, *Parathurammina brazhnikovae* Vdov., and *Radiosphaera ponderosa* Reitl. is a charac-

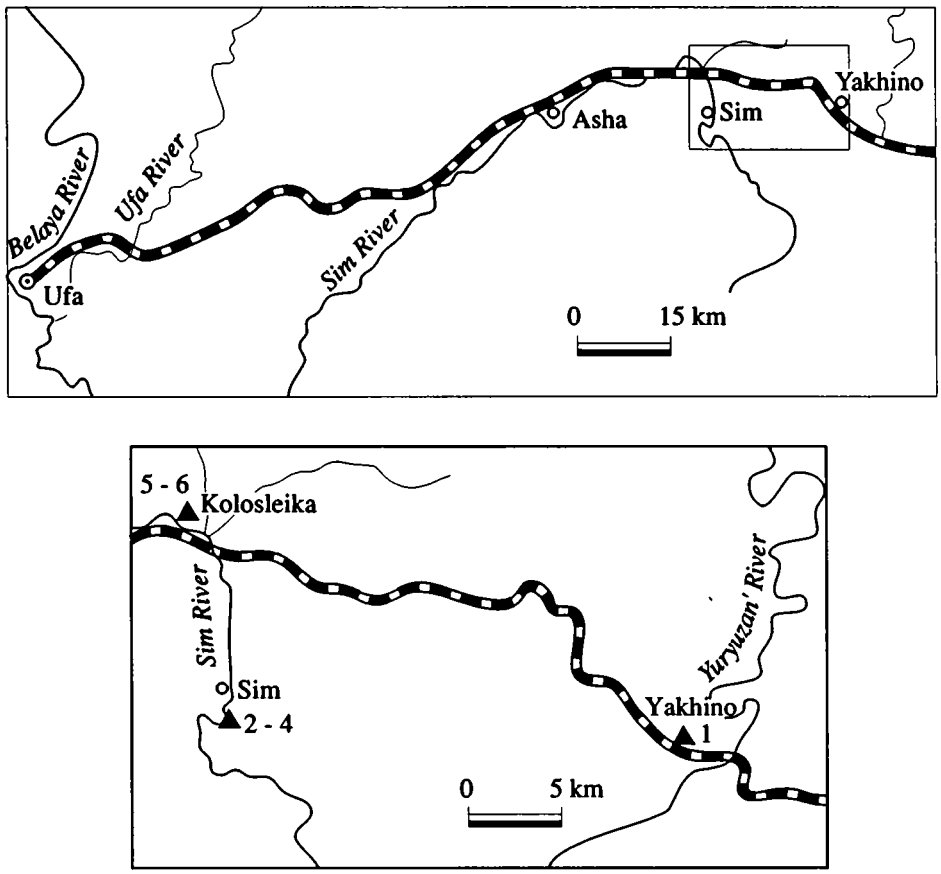


Fig. 1. *Location map of the studied sections of the Bashkirian in the Sim brachysyncline.*
1 – Yakhino section, 2 – Zhukova Shishka, 3 – outcrop 1724, the right bank of the Sim Pond; 4 – outcrop 3/38, the right bank of the Sim River, 2 km south of the Zhukova Shishka Mountain, 5 – outcrop 1729, Koloslei Village, 6 – outcrop 9/89, the right bank of the Sim River, downstream of Simskaya railway station.

teristic feature, whereas singular specimens of *Profusulinella staffellaeformis* were found only in one section (VI) near Simskaya station. In the upper layers of the horizon, first *Ozawainella pararhomboidalis* f. *minima* were found. *Donezella*, *Claracrusta* and *Masloviporidium delicata* (Berch.) have a rock-forming significance, and *Stacheoides meandriformis* Mamet et Roux and *Aoujgalia variabilis* Term. et Term. often join them in this respect. Productids, *Choristites bisulcatiformis* Semich., *Ch. uralicus asiatica* Frcks., *Ch. yanghukovensis baschkirica* Semich., *Marginifera* sp., and other brachiopods, tetracorals, *Chaetetes* aff. *capillaris* Phill., crinoids, and gastropods are distributed throughout the entire limestone sequence.

In section VI (Simskaya station) only a part of the Askynbash Horizon is exposed, but its bottom is well expressed morphologically. Thin-bedded, dark with brown tint, pelitomorphic, wavy cross-bedded limestones of the Akavas Horizon are replaced by massive, earthy-grey dolomite with *Syringopora*, solitary tetracorals, small *Productus*, and Crinoids. The thickness of the Askynbash Horizon varies between 28 and 54.5 m.

THE UPPER BASHKIRIAN SUBSTAGE

Carbonate sediments of the upper Bashkirian have been identified in the Sim brachysyncline only as the lower Tashasta Horizon. The uppermost part of the stage seems to be a constituent of the Kolosleika Formation, consisting of black claystone, cherts, pelitomorphic limestone, and marlstone with reduced thicknesses. The Kolosleika Formation also comprises the Moskovian Stage (Table 1).

The Tashasta Horizon corresponding to the *Ozawainella pararhomboidalis* foraminiferal zone and was found in sections II, III, and V; in section I the horizon was identified provisionally (Fig. 2). The boundary with the Askynbash Horizon is drawn according to the appearance of the zonal species in the section. The most complete section of the Tashasta Horizon crops out on the Zhukova Shishka Mountain (II) where the lower and upper Bashkirian limestones are separated by a deep and narrow ravine. Beyond the ravine, dark-grey and earthy-grey with brown tint, very fine-grained, bedded (bed thickness is from 5 to 23 cm), limestones are exposed, occurring almost vertically. Closer to the top, they contain nodules of light flints. Under the microscope, it can be seen that a pelitomorphic lime-

stone dominates, but there are also algal, oolitic and polydetrital limestones with small productids, *Choristites*, crinoids, and foraminifers: *Tolipammina*, *Palaeonubecularia*, large *Bradyina*, *Novella evoluta* Grozd. et Leb., *Pseudostaffella antiqua grandis* Schlyk., *Ps. krasnopolskyi* (Dutk.), *Ps. praegorskyi* Raus., *Ps. subquadrata* Grozd. et Leb., *Ps.* ex gr. *latispiralis* Kir., *Schubertella obscura* Lee et Chen, *Ozawainella aurora* Grozd. et Leb., *Oz. alchevskiensis* Pot., *Oz. pararhomboidalis* Man. f. *minima*, *Oz. pararhomboidalis* Man., and *Profusulinella parva* (Lee et Chen).

In other sections of the Sim brachysyncline where the Tashasta Horizon was defined, the following species were found too: *Eostaffella mutabilis* Raus., *E. mutabilis rjasanensis* Raus., *Schubertella mosquensis* Raus., *Sch.* ex gr. *pauciseptata* Raus., *Pseudostaffella gorskyi* (Durk.), *Ps. proozawai* f. *maxima* Sin., *Ozawainella plana* Pot., and *Profusulinella staffellaeformis* Kir. It is noteworthy that there is a somewhat increased content of red algae *Stacheoidella spissa* Petr. et Mamet, *Stacheoides meandriformis* Mamet et Roux, *St. tenuis* Petr. et Mamet, *Aoujgalia variabilis* Term. et Term., *Komia abundans* Korde, *Urtasimella* sp., and others. In addition to *Donizella*, *Masloviporidium* and *Claracrusta catenoides* (Hom.) are rock-forming species, too.

The uppermost limestone unit of the Yakhino section, 2 m thick, was tentatively included into the Tashasta Horizon on the basis of found *Pseudostaffella krasnopolskyi kyzelensis* Grozd. et Leb., which is characteristic of the Late Bashkirian and Early Moskovian (Vereya) sediments. The limestone is dark-grey to black, bedded, fine-grained, silicified, algal-foraminiferal, with the assemblage of foraminifers and algae typical of that time interval.

Of all studied sections in the Sim brachysyncline, the Tashasta deposits are thickest in the Zhukova Shishka section (27.3 m). Their thickness is quite commensurable with the thicknesses of the rest of the Bashkirian horizons in that section. *Profusulinella* are rare there. This paleontological incompleteness is compensated by abundant and diverse *Pseudostaffella* and by several species of *Schubertella* and *Ozawainella* that give an adequate idea about the Tashasta Horizon of the upper Bashkirian.

Horizon-by-horizon comparison of the microfaunal assemblages in the Sim brachysyncline and in the stratotype on the Askyn River, which are 100 km apart, but in the same structural facies zone, not only shows their similarity both in fauna and lithology but also confirms the validity of the chosen stratotype of the Bashkirian on the Askyn River. The Yakhino section in the Sim brachysyncline give a more comprehensive idea about the Bogdanovka Horizon.

Despite some differences in composition of the foraminiferal assemblages concerning mainly the number of the specimens in the taxons, all Bashkirian horizons from the stratotype and the Sim brachysyncline can be confidently correlated with the sediments of the same age from the Volga-Ural region, the Ural eastern slope, Donets Basin, Central Asia, Central Kazakhstan, that is, from all the regions in the same biogeographical belt.

Having the detailed data on the microfauna from the border sediments of the Lower and Middle Carboniferous and of the Bashkirian from the Sim brachysyncline, as well as from the stratotype on the Askyn River, one can infer some regularities in substantiating the stratigraphic boundaries at the level of series, stages, substages, and horizons.

On the western Ural slope, between the Yuryuzan'-Ai and Belaya basins, in the upper part of the Serpukhovian, we have found, for the first time, the Staroutkinsk Horizon, which is analogous to the Zapaltyubinsk Horizon on the East European Platform (Unifitsirovannye,... 1988) whose distinctive feature is the presence of few Okan *Janischewskina*, *Endothyranopsis*, and *Archaediscus*, abundant Serpukhovian *Striatifera* with *Plectostaffella*, *Eostaffella*, Archaediscidae and alga *Donezella*, which are characteristic of the Bashkirian. The identification of the Bogdanovka Horizon in the Urals is very significant for resolving the problem of the boundary between the Lower and Middle Carboniferous, which, according to the Resolution of the International Subcommission on Carboniferous Stratigraphy, should be drawn at the base of the *Homoceras* cephalopod zone. Goniatites are rare in the Urals and the search for them is difficult, therefore, the biostratigraphic basis for drawing the boundaries between the stratigraphic units and for their correlation should be foraminifers.

It should be noted that the lower boundary of the Bashkirian is very clearly visible both on the Askyn River and in the Sim brachysyncline sections. It is pronounced lithologically and faunistically, since the Bogdanovka Horizon corresponds everywhere to the member without *Striatifera* and *Choristites*. We have placed the boundary between the Lower and Middle Carboniferous at the base of the *Pl. bogdanovkensis* foraminiferal zone, which corresponds to the base of the *Homoceras* goniatite zone and to the formation of characteristic features of the conodont *Declinognathodus* genus. It is this boundary that is marked by the important changes in the phylogeny of the index foraminiferal groups (Fig. 3). For this boundary, such groups consist of the representatives of the Eostaffellidae, Biseriamminidae, and Lasiodiscidae families, and this was convincingly demonstrated by E.I. Kulagina (1990) with reference to a number of geological sections on both slopes of the South Urals.

The data we obtained from the Sim brachysyncline and the Bashkirian stratotype on the Askyn River confirm the features noted by Kulagina concerning the evolution of Eostaffellidae and Biseriamminidae at the boundary between the two, series of the Carboniferous except Lasiodiscidae because of their rarity. These features, with some additions, are as follows: consolidating the oscillation of coiling axes up to 90° for Eostaffellidae resulting in the origin of the *Plec-*

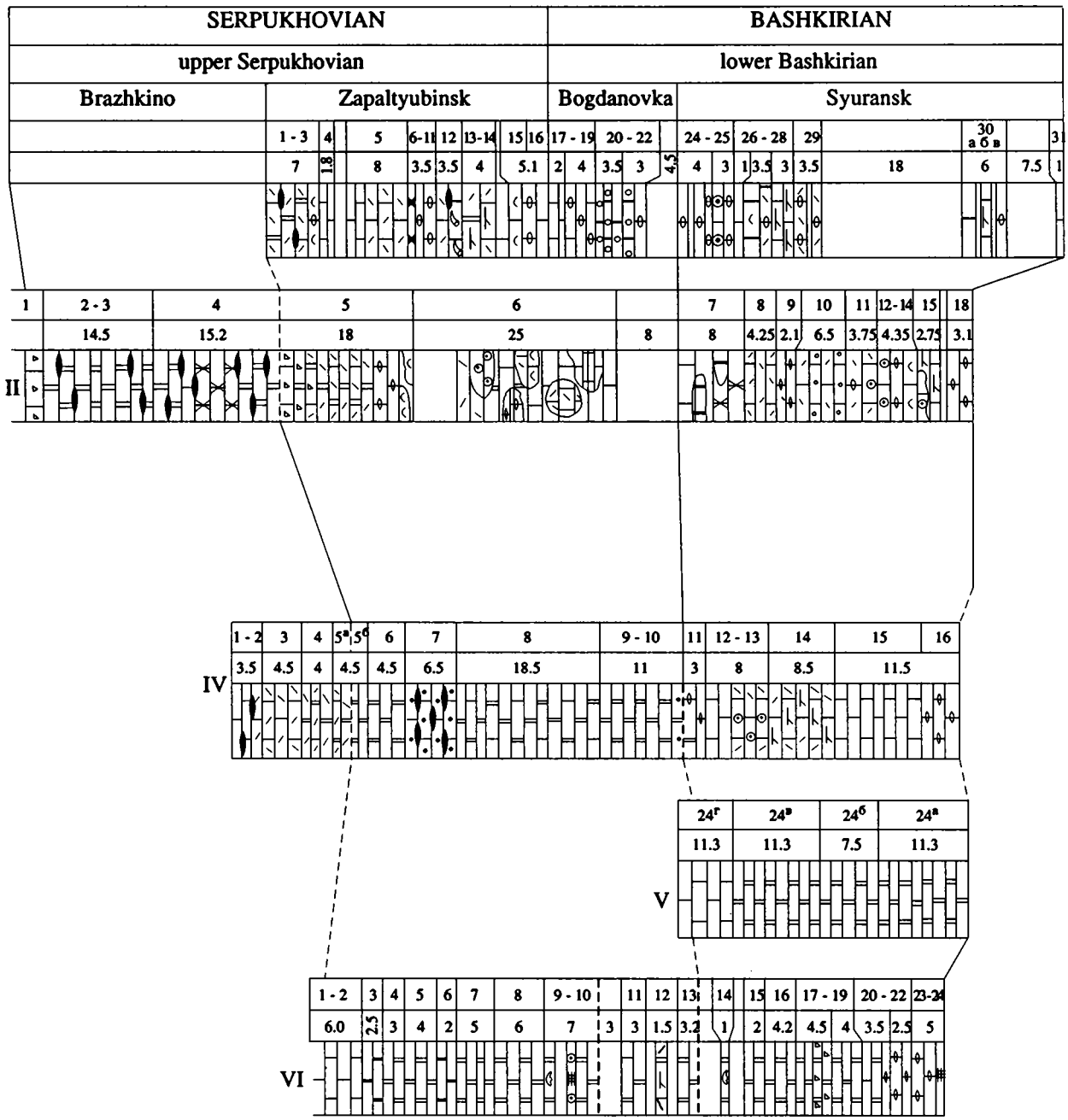


Fig. 2. *Correlation of the Bashkirian sequences in the Sim brachysyncline.*
I – Yakhino, II – Zhukova Shishka, III – outcrop 1724, IV – outcrop 3/88, V – outcrop 1729, Koloslei Village, VI – outcrop 9/89, Simskaya railway station; Limestones: 1 – pelitomorphic, 2 – polydetrital, 3 – algal, 4 – algal-polydetrital, 5 – foraminiferal, 6 – brachiopodal, 7 – crinoidal, 8 – briozoan, 9 – coralline, 10 – oolitic, 11 – organogenic sandstone, 12 – brecciated, 13 – microgrumous, 14 – silicified, 15 – with cherts, 16 – pelitomorphic, with lenses of organogenic-detrital limestones of other types, 17 – small-pebbled conglomerate, 18 – dolomite, 19 – dolomized limestone, 20 – claystone, 21 – hiatus, 22 – not exposed, 23 – presumed boundaries.

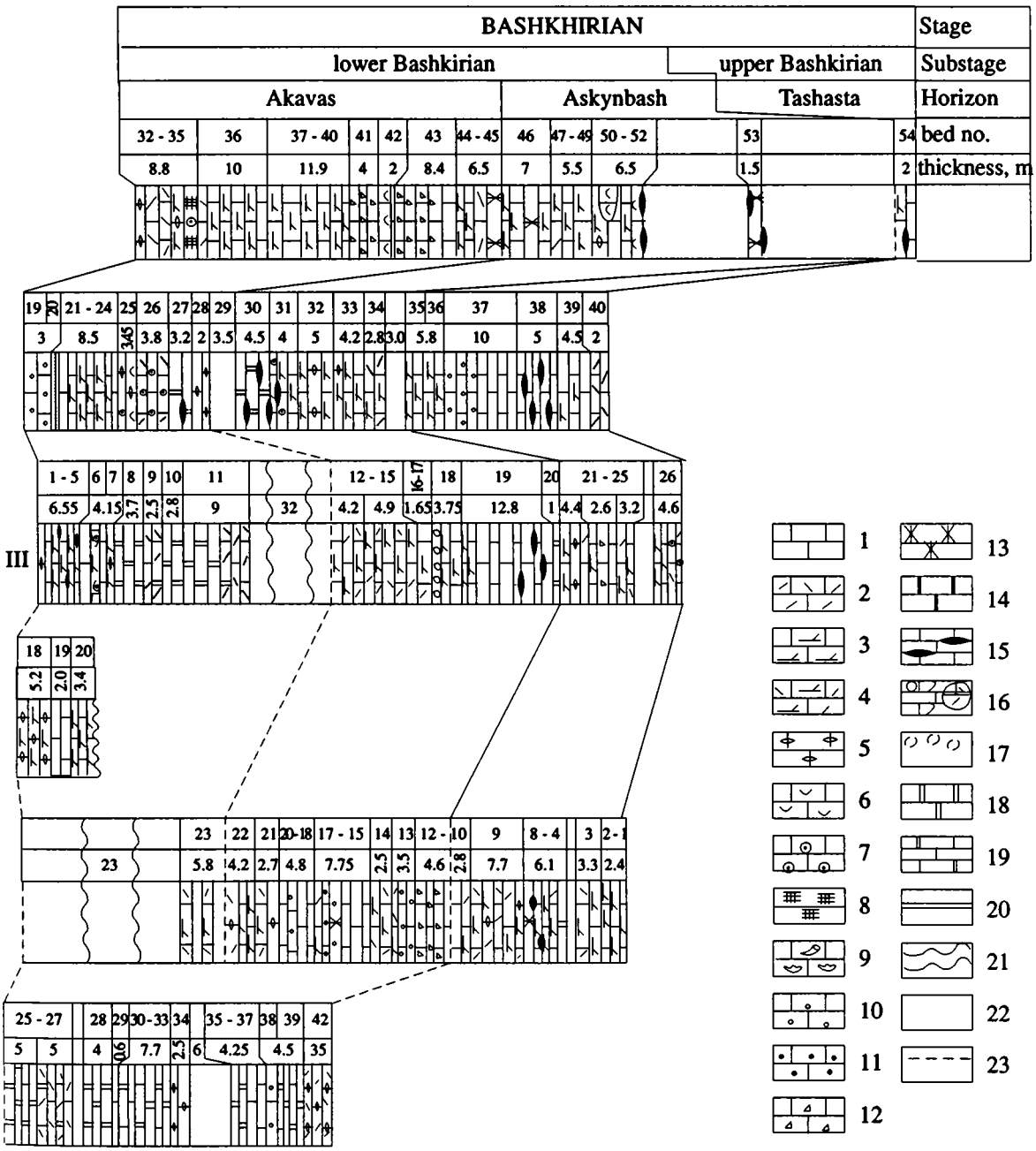

BASHKHIRIAN
Stage
lower Bashkirian
upper Bashkirian
Substage
Akavas
Askynbash
Tashasta
Horizon
32 - 35 36 37 - 40 41 42 43 44 - 45 46 47 - 49 50 - 52 53 54
bed no.
8.8 10 11.9 4 2 8.4 6.5 7 5.5 6.5 1.5 2
thickness, m
19 20 21 - 24 25 26 27 28 29 30 31 32 33 34 35 36 37 38 39 40
3 8.5 3.45 3.8 3.2 2 3.5 4.5 4 5 4.2 2.8 3.0 5.8 10 5 4.5 2
1 - 5 6 7 8 9 10 11 12 - 15 16-17 18 19 20 21 - 25 26
6.55 4.15 3.7 2.5 2.8 9 32 4.2 4.9 1.65 3.75 12.8 1 4.4 2.6 3.2 4.6
III
18 19 20
5.2 2.0 3.4
23 22 21 20-18 17 - 15 14 13 12 - 10 9 8 - 4 3 2-1
23 5.8 4.2 2.7 4.8 7.75 2.5 3.5 4.6 2.8 7.7 6.1 3.3 2.4
25 - 27 28 29 30-33 34 35 - 37 38 39 42
5 5 4 0.6 7.7 2.5 6 4.25 4.5 35
1 2 3 4 5 6 7 8 9 10 11 12
13 14 15 16 17 18 19 20 21 22 23

Fig. 2. (Contd.)

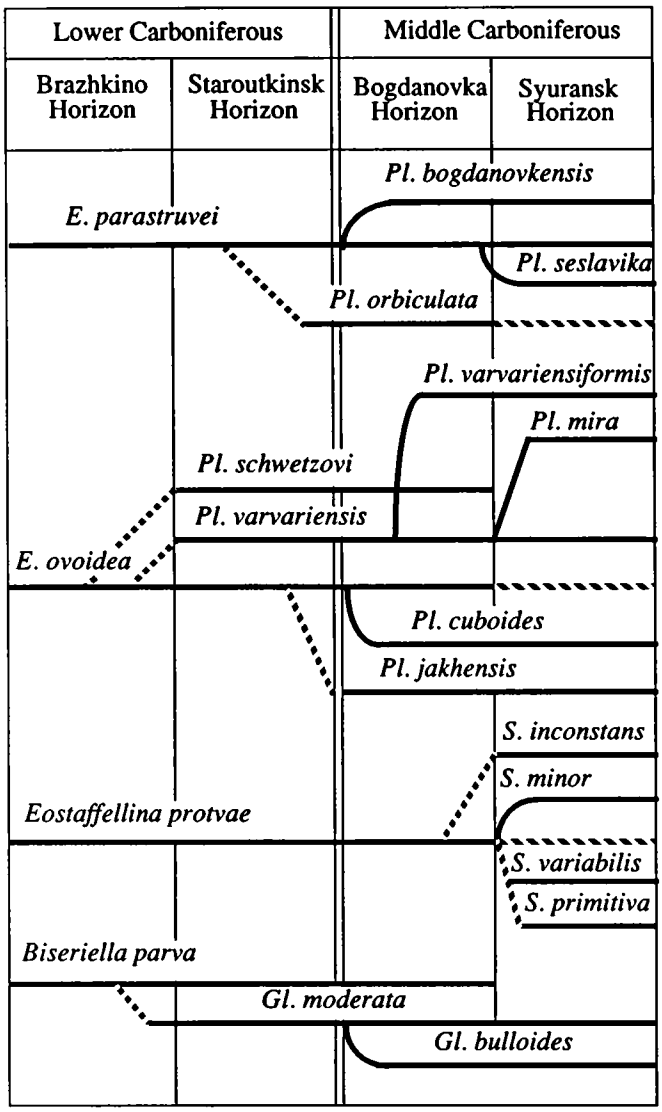


Fig. 3. *Schematics of the phylogenetic evolution of the Plectostaffella, Semistaffella, and Globivalvulina genera at the boundary of the Lower and Middle Carboniferous.*

tostaffella genus and afterwards of *Semistaffella*; formation of three-layered wall of *Biseriamminidae*. The displacement of the coiling axes has also occurred in *Millerella* and *Mediocris* genera. The intensification of the flattened *Eostaffella* evolutionary line can be observed, which resulted in the origin of the *Novella* genus in the younger horizons of the stage.

During Bogdanovka time, rapid speciation of the *Plectostaffella* genus occurred (8 species and subspecies were found in the Yakhino section), as did the evolution of the foraminifers of the *Eostaffella pseudostruvei - E. postmosquensis* line, of rather large *Globivalvulina* with thin but differentiated walls, of *Endothyra*, from the *bowmani* group, of *Semiendothyra surenica* with three-layered walls, *Planoendothyra aljutovica*, and *Endotaxis brazhnikovae*. Noticeable is a large number of attached forms, especially, *Tolipammina* and *Haplophragmina* and a complete absence of the Okan elements of the microfauna. The first primitive *Semistaffella* appear near the top of the horizon. *Plectomillerella* with displaced coiling axes are typical, but generally, the boundary between the Bogdanovka and the overlying horizon is less evident than its lower boundary. The foraminiferal assemblage is scanty in the stratotype due to dolomitization of the limestone.

The new Unified Biostratigraphic Scale of the Ural Carboniferous (1990) places the boundary between the series at the base of the goniatite H zone and of the *Pl. bogdanovkensis* foraminiferal zone, in contrast to the Unified Scale for the East European Platform (Unifitsirovannye...,1988). By identifying the Bogdanovka Horizon in the Urals, the stratigraphic range of the Syuransk Horizon was reduced by an entire zone, and now it corresponds only to the *Reticuloceras* genozone in accordance with the stratotype.

The most notable features of Syuran time are the predominance of small, elongated *Eostaffella* or "star-shaped" Archaediscidae in the biocenoses, the appearance of zonal species of *Semistaffella variabilis*, large Bradyina at the base of the horizon, and the increase in specific diversity of *Millerella* and *Plectostaffella* genera, the number of whose taxons here reaches 12. Closer to the top of the horizon, *Semistaffella* appear which have some resemblance to small *Pseudostaffella*. Some recurrent forms (*Bituberitina* sp. and *Paracaligella* sp.) and elongated *Millerella* are present. Among them there are individuals resembling to *M. marblensis* and characteristic of the Pennsylvanian of North America. In all sections where the SyuranHorizon was identified, its lower boundary is marked by the appearance of *Semistaffella variabilis*. It should also be noted that the foraminiferal assemblages from the Bogdanovka and Syuran horizons are close to one another and difficult to differentiate, so in a number of sections in the Urals the lower parts of the sequences attributed to the Syuransk Horizon seem to unite the two goniatite zones: *Homoceras* and *Reticuloceras*. This was taken into consideration in the new Unified Scale of the Ural Carboniferous (1990). Mass accumulations of Martinia, forming a coquina interbed, are also the characteristic feature of Syuransk time.

The lower boundary of the Akavas Horizon can be unambiguously identified by the first appearance of *Pseudostaffella antiqua* in the sequence, however, some exceptions to the rule were observed in several sections of the Sim brachysyncline, as well as on the eastern slope of the Urals. Sometimes, due to specific facies features of section, the index species is absent at the base of the horizon and even in its lower half, and first *Pseudostaffella* are hardly distinguishable from *Semistaffella*. This concerns, in particular, such species as *Ps. ziganica*, *Ps. korobezkikh*, and *Ps. varsanofievae*, which also mark the beginning of Akavas time when the zonal species is absent. In these cases an important role in the paleocommunities is played by *Eostaffella* with elongated keeled peripheral part, various *Parastaffella*, and *Semistaffella*. Larger *Pseudostaffella* with

clear generic features appear a little later, approximately at the base of the second third of the horizon. These forms, *Ps. antiqua grandis* and *Ps. antiqua posterior*, gradually supplant *Semistaffella*. Rare *Schubertella* and small rhombic *Ozawainella* can be observed in the second half of the horizon, but their further evolution occurs later.

The blossom of *Pseudostaffella* and appearance of the group of rather large forms among them, with flattened peripheral part and chomata thicker than that of the Akavas species, is related to Askynbash time. These are represented, along with the zonal species *Ps. praegorskyi*, by *Ps. krasnopolskyi*, *Ps. composita*, *Ps.* ex. gr. *nibelensis*, and others. *Pseudostaffella* from the group of *Ps. ozawai*, elongated with convex side slopes (*Ps. proozawai*, *Ps. proozawai acuta*, and *Ps. proozawai* f. *unbolata*, minima, maxima), occurred at the same time. Forms with sharpened periphery and flattened side slopes, resembling the Moskovian species *Ps. topilini*, are frequent. The appearance of these *Pseudostaffela* is an important criterion for the definition of the boundary between the Akavas and Askynbash horizons. Such a feature of Askynbash *Pseudostaffella* can be traced throughout the entire Ural region. The significance of *Semistaffella* and *Plectostaffella* here is reduced to zero. First spherical *Profusulinella staffellaeformis* appear simultaneously with *Ps. praegorskyi*, and singular *Profusulinella* ex. gr. *parva*, only at the end of the Askynbash time and not in each sequence. The significance of *Schubertella* and small rhombic *Ozawainella* becomes much more noticeable than in Akavas time. It should be noted that such *Ozawainella* later appear in the East European province.

The development of *Ozawainella*, *Profusulinella*, *Schubertella*, and *Pseudostaffella* with a more complex organization than in the early Bashkirian, is characteristic of the late Bashkirian in the Sim brachysyncline and in the stratotype. First *Aljutovella* are known only on the Askyn River because the uppermost Asatau Horizon of the Bashkirian is represented in the Sim brachysyncline by the Kolosleika Formation which does not contain foraminifers.

The progress of rhombic *Ozawainella* is characteristic of Tashasta time everywhere. They usually have a convex umbilical area and, in contrast to the early Bashkirian specimens, a larger size, a well-developed keel, and rather thick, ribbon-like chomata. Unfortunately, a zonal species of *Ozawainella pararhomboidalis* is not present everywhere. *Profusulinella parva* and *Pr.* ex. gr. *parva* occur in the assemblage almost constantly. The specific composition of *Pseudostaffella* changes markedly towards a decrease in taxonomic diversity, and the rejuvenation and increase in size of shells and chomata. *Ps. gorskyi*, *Ps. subquadrata*, *Ps.* ex gr. *latispiralis*, and *Ps. krasnopolskyi kyzelensis* typical of the Early Moskovian sediment start to be met along with *Ps. praegorskyi*. The forms from a *Ps. ozawai* group decrease in abundance, *Schubertella* become more diverse and first *Novella* are noted. The characteristic feature of the sections we studied in the Sim brachysyncline and of the stratotype, is rare occurrence of *Profusulinella* and quantitative dominance of *Ozawainella* and *Pseudostaffella*, as well as the absence of the beds with *Profusulinella primitiva* that were distinguished by G.D. Kireeva (1971) along the Zilim River and are characteristic of other regions. We did not find *Pr. rhomboformis* and *Pr. oblonga* either, which she indicated for the lower part of the Tashasta Horizon, except for one questionable fragment of *Pr.* ex. gr. *rhombiformis* in the Askyn section (Sinitsyna and Sinitsyn, 1987). This situation seems to be related to unfavorable environmental conditions for *Profusulinella*, because the sediments of the Tashasta Horizon consist mainly of algal, slime, and pelitomoraphic limestones.

The uppermost Asatau Horizon is known only in the stratotype where rare *Aljutovella tikhonovichi* and *Profusulinella* likewise rare, but more diverse in composition than those of the Tashasta Horizon were found. Singular *Profusulinella prisca* and *Pr.* ex gr. *prisca* were added to the already known species. The absence of *Verella* in the stratotype, and the scarcity of *Profusulinella* in the upper Bashkirian testify that the conditions for their evolution were unfavorable because the main part of all studied sections was occupied by algae, especially red algae and green *Donezella* and *Claracrusta*. Numerous *Verella* appear farther north, in the Middle Urals in the Chusovaya River sections surrounding the northern end of the Ufa Amphitheater.

CONCLUSION

The Staroutkinsk Horizon (the upper one of the Serpukhovian Stage) and Bogdanovka Horizon (the lower one of the Bashkirian stage) were first defined as the result of examination of the deposits at the boundary between the Lower and Middle Carboniferous. The data obtained confirm the absence of a hiatus at this boundary.

The Bashkirian was divided into two substages that were divided in turn into horizons and foraminiferal zones. The upper Bashkirian sediments were also first identified in the Sim brachysyncline.

All the foraminiferal zones defined in the studied sections of the brachysyncline are correlatable with zones of the stratotype at the Askyn River, clarify their characteristics, and enhance the correlating possibilities. The foraminiferal biostratigraphic assemblages are liable to change considerably under the influence of environmental conditions; nevertheless, their "main body" can be found in the territory of the Urals, East European Platform, Donets Basin, Tien Shan and Central Kazakhstan.

The algal communities were identified for the first time too, and they allow appreciation of the foraminiferal assemblages in the context of their dependence on facies.

REFERENCES

Chuvashov, B.I. and Mizens, G.A., *Verkhnii Paleozoi R. Ai (Zapadnyi Sklon Yuzhnogo Urala)* (The Upper Paleozoic of the Ai River Region (the Western Slope of the South Urals)), IGiG, UrO AN SSSR, Sverdlovsk, 1991, pp. 17 - 23.

Dmitriev, G.A. and Nalivkin, V.D., The Upper Paleozoic of the Sim Factory, *Problemy Sovetskoi Geologii*, 1927, no. 7, pp. 607 - 616.

Einor, O.L., *Issledovaniya po Stratigrafiyi Karbona Vostochnoi Okrainy Uralo-Volzhskoi Neftenosnoi Oblasti (Gornaya Bashkiriya)* (Investigations on the Carboniferous Stratigraphy in the Eastern Margin of the Volga-Ural Petroliferous Province (Mountainous Bashkiria)), Gostoptekhizdat, Moscow, 1958.

Einor, O.L., Furdui, R.S., Alexandrov, V.A., *et al.*, The Bogdanovka and Syuransk Horizons of the Carboniferous in the Bol'shoi Suren' River Basin (Mountainous Bashkiria), *Doklady AN SSSR*, 1973, vol. 213, no. 5, pp. 1155 - 1157.

Groves, J.R., Calcareous Foraminifers from the Bashkirian Stratotype (Middle Carboniferous, South Urals) and Their Significance for Intercontinental Correlations and the Evolution of the Fusulinidae, *Journal of Paleontology*, 1988, vol. 62, no. 3, pp. 368 - 399.

Kireeva, G.D., More Precise Definition of the Bashkirian in the Mountainous Bashkiria on the Basis of Studying Foraminifers, Biostratigrafiya Otlozhenii Bashkirskogo Yarusa Volgo-Ural'skoi Oblasti, *Trudy VNIGNI*, 1971, no. 107, pp. 6 - 54.

Kulagina, E.I., *Biostratigrafiya pogranichnykh otlozhenii nizhnego i srednego karbona Yuzhnogo Urala po fo- raminiferam* (Foraminifera-Based Biostratigraphy of the Sediments at the Boundary Between the Lower and Middle Carboniferous of the South Urals), Cand. Sci. (Geol.-Min.) Dissertation, Sverdlovsk, 1990.

Malakhova, N.P. and Malakhov, A.A., The Moskovian and Gzhelian Stages of the Middle Urals, *Trudy Gorno-Geologicheskogo Instituta, Ural'skii Filial, AN SSSR*, 1961, no. 59, p. 161.

Petrenko, A.A. and Kiparisova, L.D., Notes on Stratigraphy of the Carboniferous Deposits from the Western Slope of the South Urals. I. Stratigraphy of the Carboniferous Deposits Near Yakh'ya Village, *Bull. MOIP, Otd. Geol.*, 1937, vol. 15 (6), no. 2, pp. 541 - 563.

Reitlinger, E.A., On the Boundary Between the Bogdanovka and Krasnopolyansk Horizons (Foraminifers of the *Homoceras* Zone), *Voprosy Mikropaleontologii*, 1980, no. 23, pp. 23 - 38.

Sinitsyna, Z.A. and Sinitsyn, I.I., *Biostratigrafiya Bashkirskogo Yarusa v Stratotipe* (Biostratigraphy of the Bashkirian in the Stratotype), Bashkirskii Filial AN SSSR, Ufa, 1987.

Unifitsirovannye i Korrelyatsionnye Skhemy Devona i Karbona Russkoi Platformy (Unified and Correlative Schemes of the Devonian and Carboniferous of the Russian Platform), VSEGEI, Moscow, 1988.

Stratigraphy and Geological Correlation, Vol. 1, No. 1, 1993. Translated from Stratigraphiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 1, 1993, pp. 87 - 96.
Original Russian Edition Copyright © 1993 by Stratigraphiya. Geologicheskaya Korrelyatsiya, Herman.
English Translation Copyright © 1993 by Interperiodica.

Evolution of the biosphere

Stages and Cycles in the Late Cretaceous Floral Changes of the Anadyr'-Koryak Subregion (Northeast Russia) and Their Connection with Climatic Changes

A. B. Herman

Geological Institute, Russian Academy of Sciences, Pyzhevskii Per. 7, Moscow, 109017 Russia

Manuscript received March 20, 1992

Abstract – The Cretaceous continental flora-bearing sediments of the Anadyr'-Koryak Subregion are dated on the basis of their alterations with marine layers. Taphofloras from eleven sections of the subregion are classified into six evolutionary stages within the Late Cretaceous: Grebenka (latest Albian ? to early Turonian), Penzhina (Turonian, probably excluding its beginning), Kaivayam (Coniacian), Barykov (Santonian to early Campanian), Gornaya (middle Maestrichtian), and Koryak (Late Maestrichtian), with a hiatus in the late Campanian and early Maestrichtian. Based on these floras, warming in the Cenomanian and cooling in the Turonian and Coniacian followed by a warming in the Santonian and early Campanian, and a cooling by the end of the Cretaceous have been reconstructed. The inferred Late Cretaceous climatic cycles (warming + cooling) were the likely cause of the cyclic pattern of paleofloral changes. Similar paleoclimatic variations are noted for other regions of the world and probably have a global significance.

Northeast Russia is divided into three subregions differing in their geological history (Fig. 1a): the Verkhoyansk-Chukotka, Okhotsk-Chukotka (a volcanic belt), and Anadyr'-Koryak. The first features of Cretaceous sedimentation which occurred in isolated coal-bearing basins, and the second, thick terrestrial volcanic accumulations. In the Upper Cretaceous in the Anadyr'-Koryak Subregion (AKSR) continental flora-bearing sediments were accumulated in the coastal plains along with marine sediments containing ammonite and inoceram fossils. Very often, while observing the interrelation between these deposits, it was possible to correlate the fossil plant localities with the marine stratigraphic scale.

The analysis of systematic composition and stratigraphic position of the taphofloras from eleven flora-bearing sections of AKSR (Fig. 1b) allows to identify six stages in the Late Cretaceous flora evolution in this region (Fig. 2). The flora-bearing AKSR sediments are correlated and phytostratigraphic horizons identified on the basis of the periodicity of this flora evolution (Fig. 3). To ensure a stable nomenclature of the stages and of the corresponding horizons, for each of them the type floras are indicated; the subsections containing the type floras are regarded as stratotypes of the phytostratigraphic horizons.

The Grebenka Stage
(Late Albian ? - Cenomanian - ? Early Turonian)

This stage includes the Grebenka flora (comprising a number of taphofloras) from the Anadyr' River middle reaches (Shchepetov, Herman, 1990; Shchepetov *et al.*, 1992), and the middle Ginter flora from the Ugol'naya Bay (Herman, 1988). The taphofloral assemblage from the Krivaya Formation in the Grebenka River middle reaches is the richest example of Grebenka flora and is considered the type flora for this stage.

The following main features can be outlined for taphofloras of the Grebenka stage. Angiosperms are broadly diversified and comprise about 35% of the total quantity of species. Typical is the high abundance of representatives of the *Menispermites* genus: *M. ginterensis* Herman, *M. marcovoensis* Philippova, *M. minimus* (Krysht.) Herman, comb. nov., *M.* ex gr. *septentrionalis* Hollick, large-leaved Platanoids, which are not yet significantly diverse, are appreciable although not predominant. Lobed leaves of *Araliaephyllum* and compound leaves referred to *Schefflerаephyllum*, but in fact probably belonging to the *Debeya* genus, are common. Representatives of the *Magnoliaephyllum*, *Dalbergites*, *Celastrophyllum*, *Cissites*, *Dalembia*, *Grebenkia*, *Sorbites*, *Leguminosites*, and other genera have also been found. Findings of *Trochodendroides* are rare. Among ferns, *Gleichenites*, *Birisia*, and *Coniopte-*

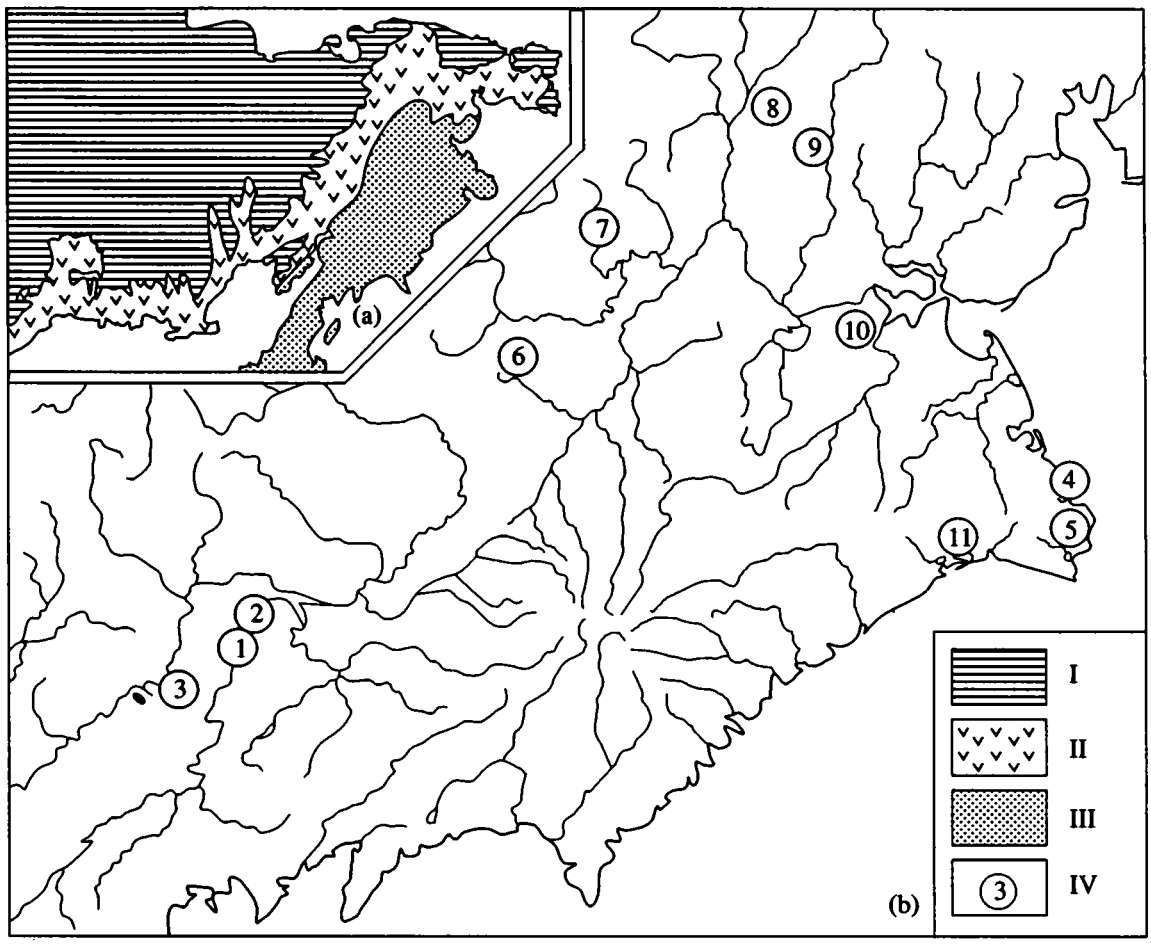


Fig. 1. *Regions of Northeastern Russia (a) and Late Cretaceous flora locations of the Anadyr'-Koryak Subregion (b).*
I – Verkhoyansk-Chukotka subregion; II – Okhotsk-Chukotka subregion (a volcanic belt); III – Anadyr'-Koryak subregion, IV – Main localities of fossil flora (numbers on the scheme): 1 – Cape Konglomeratovy, 2 – Cape Valizhgen, 3 – Elistratov peninsula, 4 – Ugol'naya Bay, 5 – Amaam Lagoon, 6 – Grebenka River, 7 – Left shore of the Anadyr' River (Ubienka river), 8, 9 – Pekul'nei Ridge: 8 – west slope, 9 – east slope, 10 – Rarytkin Ridge, 11 – Pekul'nei Lake.

rare. Among ferns, *Gleichenites*, *Birisia*, and *Coniopteris* are common. Cycadophytes represented by *Cycadites*, *Nilssonia* (their leaves often form accumulations on bedding planes), *Nilssoniocladus*, *Taeniopteris* are obligatory components of the flora. Ginkgoales include the *Ginkgo*, *Baiera*, and *Sphenobaiera* genera. Czekanowskiales feature the genus *Phoenicopsis*, of which findings are extremely rare. Conifers highlight both relatively old (*Podozamites*, *Athrotaxopsis*, *Pagiophyllum*, *Pityophyllum*), and younger (*Sequoia*, *Thuja*) genera. The most common species among conifers are *Cephalotaxopsis intermedia* Hollick, *Araucarites anadyrensis* Krysht., and *Elatocladus smittiana* (Heer) Seward.

Penzhina Stage (Turonian, Probably, Excluding Its Beginning)

The Penzhina taphoflora from the deposits of the Valizhgen Formation in northwest Kamchatka and the Elistratov Peninsula (Herman, 1991), as well as the Volch'ya taphoflora from the basin of the Ubienka River on the Anadyr' left banks (Devyatilova *et al.*, 1980), correspond to this stage. The Penzhina taphoflora from Cape Konglomeratovyi (northwest Kamchatka) is the type flora for this stage.

The following features are inherent in the Penzhina taphofloras. Large-leaved platanoids have come to the fore in diversity and frequency of occurrence. They are represented by old species of *Platanus*, *as well as by genera Arthollia*, and *Paraprotophyllum*. The role of *Menispermites* has somewhat declined. However, the species of this genus are common in floras of this stage (*M. kryshtofovichii* Vachr. and *M. septentrionalis* Hollick). The *Celastrophyllum* genus continues to occur and its representatives may play an important role in the composition of the floral complexes. The species common for the taphofloras of the Grebenka stage (*Gleichenites zippei* (Corda) Seward, *Menispermites septen-*

trionalis Holl., *Grebenkia kryshtofovichii* Lebed., *Dalembia pertrionalis* Holl., *Grebenkia kryshtofovichii* Lebed., *Dalembia pergamentii* Herman et Lebed., etc.) are concomitant with angiosperms characteristic of the younger floras from Northeast Asia, i.e., *Paraprotophyllum ignatianum* (Krysht. et Baik.) Herman, *Viburniphyllum whymperi* (Heer) Herman, "*Ziziphus*" *smilacifolia* Budantsev. The *Trochodendroides* genus is common. Its representatives are reported in high frequencies from the taphofloras of the Penzhina stage. The diversity of ferns decreases in comparison to the Grebenka stage. They are assigned to the genera *Gleichenites*, *Onychiopsis*, *Cladophlebis*, *Arctopteris*, etc. Conifers are abundant, featuring frequently occurring *Cephalotaxopsis* and *Sequoia tenuifolia* (Schmalh.) Sveshn. et Budantsev. Appearances of *Protophyllocladus*, *Metasequoia*, *Glyptostrobus* and *Pseudolarix* have been reported. Cycadophytes are conspicuously absent in the taphofloras of the Penzhina stage.

Kaivayam Stage (Coniacian)

The Kaivayam taphoflora from the Valizhgen Formation of northwest Kamchatka and Elistratov Peninsula (Herman, 1991), as well as the Poperechnaya and Tylpegyrgynai taphofloras from the Pekul'nei Ridge (Terekhova, Filippova, 1983, 1984), are assigned to this stage. The Kaivayam taphoflora from the Konglomeratovyi Cape (northwest Kamchatka) is the type flora for this stage.

The floral complexes of the Kaivayam stage have the following common features. Large-leaved platanoids continue to dominate. This particularly applies to *Paraprotophyllum*. The taphofloras of the Pekul'ney Ridge also contain *Pseudoprotophyllum*. *Trochodendroides* are numerous and diverse (3 to 5 species). *Paraprotophyllum pseudopeltatum* Herman and *Trochodendroides sachalinensis* (Krysht.) Krysht. appear. The common plants for the taphofloras of this stage are *Magnoliaephyllum magnificum* (Dawson) Krassil., *Viburniphyllum whymperi* (Heer) Herman (or close species) as well as representatives of the *Araliaephyllum*, *Magnoliaephyllum*, "*Ziziphus*," *Cissites*, *Viburniphyllum* and *Dalembia* genera. In comparison with the flora of the Penzhina stage, *Menispermites*, *Platanus*, and *Celastrophyllum* leaves are rare. Ferns are not very diverse and are referred to the *Osmunda*, *Ruffordia*, *Arctopteris*, *Birisia*, *Cladophlebis* and other genera. Ginkgoales are represented only by *Ginkgo* ex gr. *adiantoides* (Unger) Heer, while Czekanowskiales have not been recorded. Cycadophytes are either few or absent. Finds of relict mesophytic plant *Ctenis* are probably the youngest representatives of this genus in Northeast Asia. Conifers are dominated by *Sequoia* and *Cephalotaxopsis* which usually accompany *Elatocladus*, *Metasequoia*, *Glyptostrobus*, and *Thuja*.

Barykov Stage (Santonian to Early Campanian)

The Valizhgen and upper Bystraya taphofloras from the upper parts of the Valizhgen and Bystraya formations of northwest Kamchatka and the Barykov taphoflora from the upper coal-bearing beds of the Barykov Formation of the Ugol'naya Bay pertain to this stage (Herman, 1991). The latter taphoflora is regarded as type one for this stage.

The taphofloras of the Barykov stage are characterized by the following features. The predominant angiosperms are *Macclintokia* (several species, of which *M. ochotica* Vachr. et Herman is the most widespread), and "*Quercus*" *tchucotica* Abram. (the latter, however, has not been encountered in the Valizhgen taphoflora of Kamchatka). Another prominent feature of these taphofloras is the permanent presence of cycadophytes. Among them the leaves of *Nilssonia* occur very often. *Cycadites*, *Pterophyllum* and *Encephalartopsis* are less common. Representatives of the *Grewiopsis*, *Rhamnites*, "*Vitis*," and *Hollickia* genera have been observed. Of platanoids, the most widespread is *Paraprotophyllum*, but by and large these plants are already less significant than in the Turonian and Coniacian. *Magnoliaephyllum*, "*Ziziphus*," *Viburniphyllum* (especially in Kamchatka taphofloras), *Cissites* and *Quereuxia* are common for floras of the Barykov stage. *Trochodendroides* are neither numerous nor diverse. Ferns are few. Ginkgoales feature only *Ginkgo* ex gr. *adiantoides*, which may be very abundant in certain burials. Conifers are not very diverse, *Sequoia* and *Cephalotaxopsis* being the most widespread. *Metasequoia* and *Thuja* are also common. Advanced Cenophytic plants are often accompanied by relics (*Hausmannia*, *Arctopteris*, cycadophites).

It is not improbable that in the future the Barykov stage may be divided into two (or several) lesser stages. However, the presently available data, primarily on Santonian floras of AKSR, are not sufficient for this.

Gornaya Stage (Middle Maestrichtian)

This stage in the evolution of the Late Cretaceous flora of AKSR has been recently identified by L.B. Golovneva (1990). The Gornaya taphoflora from the lower half of the Rarytkin Formation of the Rarytkin Ridge and Kakanaut taphoflora from continental deposits in the vicinity of Pekul'nei Lake correspond to this stage (Golovneva, 1990; Nesov, Golovneva, 1990). The Gornaya flora is considered to be the type flora.

The flora of the Gornaya stage is characterized by the involvement of large-leaved platanoids *Platanus* and *Artholia*, predominance of *Peculnea*, *Trochodendroides*, *Celastrinites*, *Cissites*, *Liriophyllum*, and *Quereuxia* among angiosperms and appearance of *Corylus*. *Equisetum arcticum* Heer is common, while ferns represented by *Osmunda* sp., are very rare. *Ginkgo* and cycadophytes *Nilssonia* and *Encephalartopsis*, presumed to have formed monodominant assemblages,

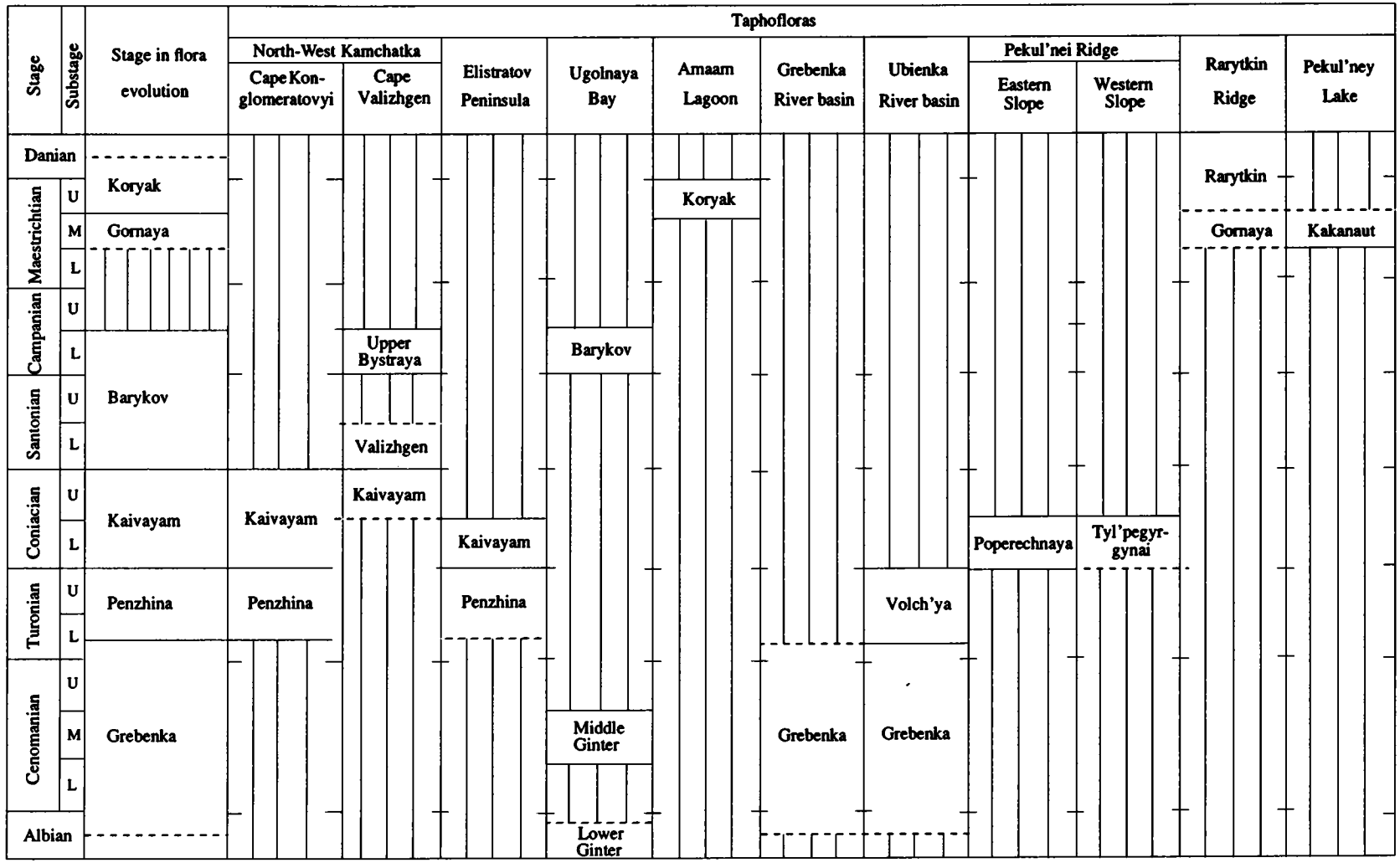


Fig. 2. *Stages of Late Cretaceous flora evolution in AKSR and main corresponding taphofloras.*

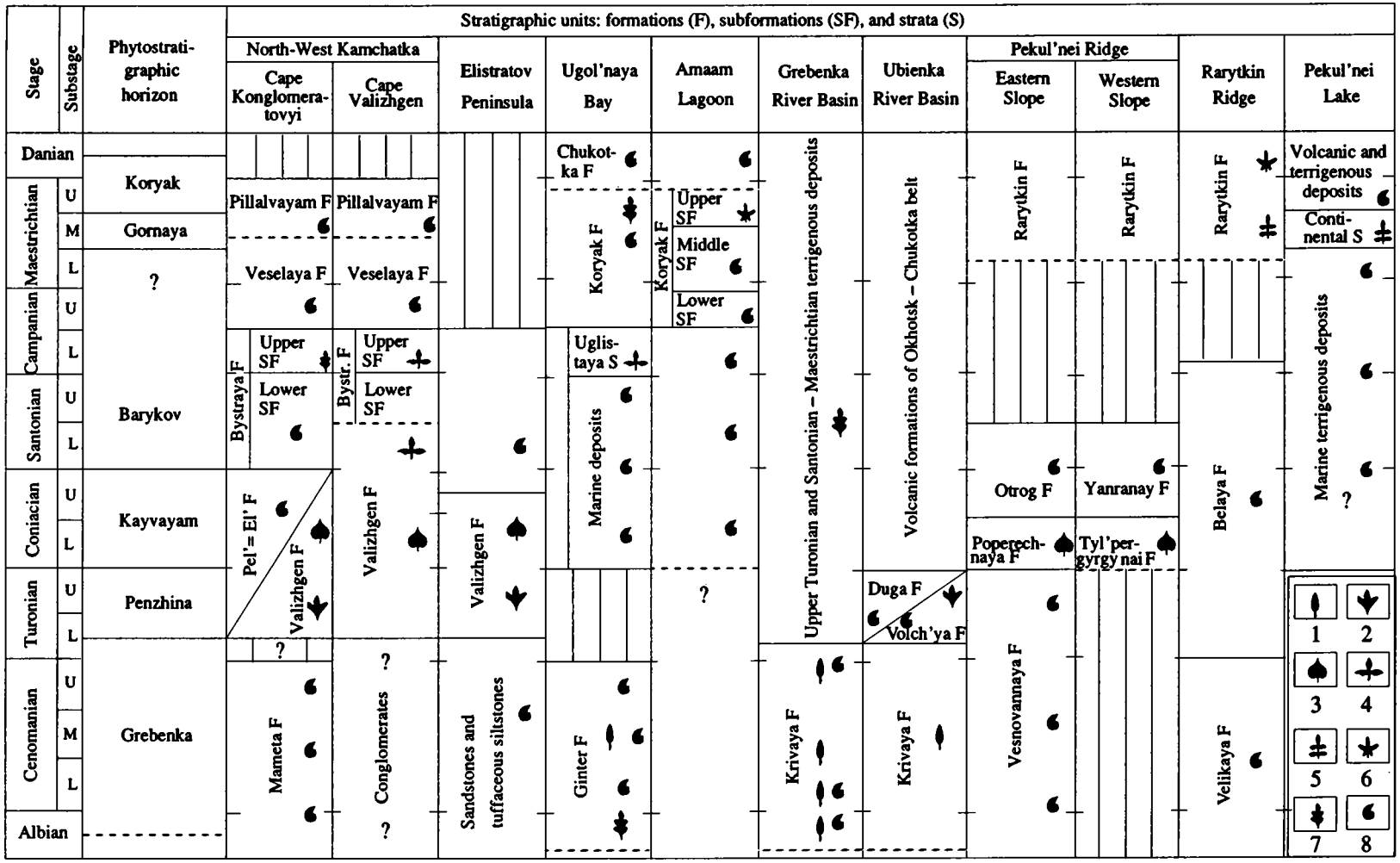


Fig. 3. *Correlation of the Upper Cretaceous flora-bearing AKSR sediments (phytostratigraphic horizons):*
1 - 7 – taphofloras of: 1 – Grebenka type, 2 – Penzhina type, 3 – Kaivayam type, 4 – Barykov type, 5 – Gornaya type, 6 – Koryak type, 7 – undetermined type, 8 – fossil marine mollusk finds, controlling the age of stratigraphic units.

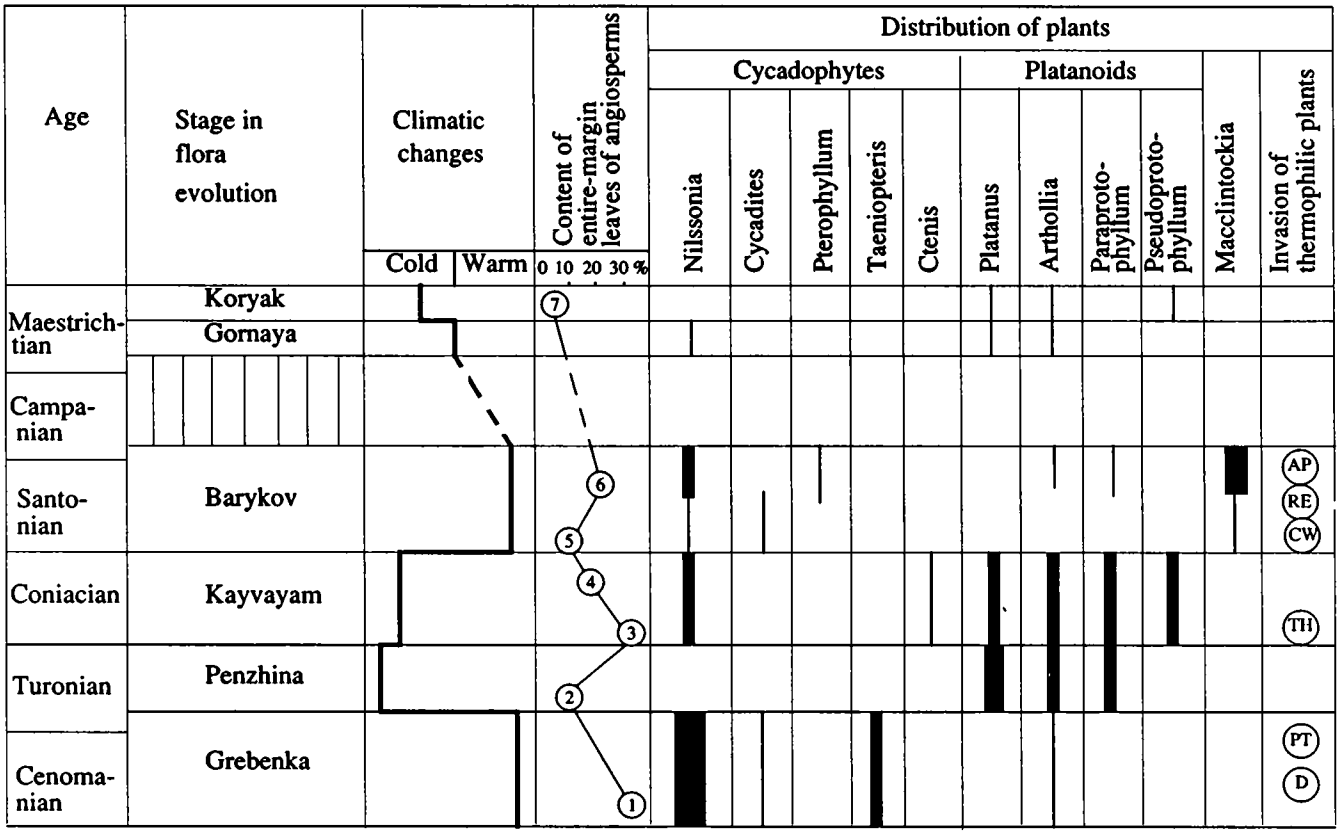


Fig. 4. *Climatic reconstructions according to paleofloristic changes in the Late Cretaceous of the AKSR. Taphofloras are marked by figures in circles:*

1 – Grebenka, the Grebenka River, 2 – Penzhina, Northwestern Kamchatka, 3 – Early Kaivayam, Northwestern Kamchatka, 4 – Late Kaivayam, Northwestern Kamchatka, 5 – Valizhgen, Northwestern Kamchatka, 6 – Barykov, Ugol'naya Bay, 7 – Koryak, interfluve of the Emima and Il'naivaam rivers. The thickness of the plant distribution lines is proportional to the quantity of species (one, two, or three) in the genus. The appearance of the plants common for southern regions is marked by letters: AP –Araliaephyllum polevoii (Krysht.) Krassil., CW – Celastrinites wardii (Knowlton) Bell, D – Debeya, PT– Pagiophyllum triangulare Pryn., RE – Rhamnites eminens (Dawson) Bell, TH – Ternstroemites harwoodensis (Dawson) Bell.

are reported in high frequencies. Conifers are not numerous, being dominated by the Late Cretaceous plants *Taxites*, *Elatocladus smittiana* (Heer) Seward and *Sequoia minuta* Sveshn. These are concomitant with younger elements, i.e., *Metasequoia*, *Glyptostrobus*, *Microconium*.

According to L.B. Golovneva (1990), the difference between the floras of the Gornaya and the Barykov stages concerns primarily angiosperms, whose composition has radically changed. At the same time, the composition of conifers has remained virtually unaltered.

The floras of the Barykov and Gornaya stages are separated by a time interval encompassing the second half of the Campanian and early Maestrichtian (7 - 8 Ma). Taphofloras of this age are not known from the territory of AKSR. Therefore, it is not clear whether there was any flora of an intermediate type at this time or the Gornaya stage of flora evolution follows directly the Barykov stage, with the boundary lying somewhere within this interval.

Koryak Stage (Late Maestrichtian)

I assigned the Koryak flora from the upper Koryak Subformation of the Amaam Lagoon and the early Rarytkin subcomplex of Rarytkin Ridge to the Koryak stage (Herman, 1989, 1990b, 1991). According to L.B. Golovneva (1990), this stage (which she calls Rarytkin) continues well into the Danian. She considers that in addition to the taphofloras enumerated, the late Rarytkin subcomplex of Rarytkin Ridge also relates to this stage. Unfortunately, there are no analogs of this subcomplex among other AKSR taphofloras. Its Danian age is determined rather provisionally, as there are no marine interbeds in the Rarytkin Formation. For this reason, I continue to consider the Koryak stage of flora evolution in its previous range until additional evidence on the Maestrichtian-Danian taphofloras of AKSR is obtained. The Koryak taphoflora from Amaam Lagoon is the type flora for this stage.

The taphofloras of this stage combine the following characteristics. The predominant plants are *Metase-*

quoia, *Corylus beringiana* (Krysht.) Golovn., *Trochodendroides* (3 - 4 species), and, occasionally, *Glyptostrobus*, *Microconium*, *Platanus reynoldsii* Newb. Less numerous are *Onoclea hesperia* R. W. Brown, *Coniopteris*, *Sequoia*, *Haemanthophyllum cordatum* Golovn., *Arthollia*, *Rarytkinia*, *Cissites*, *Celastrinites*, "*Pterospermites*" and *Nyssa*. In comparison with the Gornaya stage, angiosperms reveal increasing quantities of large-leaved plants (*Platanus*, *Arthollia*, *Pseudoprotophyllum*, *Corylus*). Conifers, represented largely by modern genera, play a major role, often being codominants with angiosperms. Equisetales and ferns are not numerous, fossils of *Ginkgo* are very rare, and cycadophytes are absent.

Together with lithological and geochemical data, paleontological and especially paleobotanical data are widely used for paleoclimatic reconstructions (Vakhrameev, 1964, 1988; Krasilov, 1975, 1979, 1985; Wolfe, 1978, 1987; Wolfe, Upchurch, 1987). Climatic changes appear to have predetermined the periodicity of the AKSR flora evolution. Due to the climatic nature of paleofloral changes in the Late Cretaceous of AKSR climatic alterations (coolings and warmings) are reflected in the composition of the flora. The following empirical criteria may, presumably, be used for its reconstruction (Fig. 4).

1. Good correlation between the leaf margin of angiosperms and climate is known: intact margins increase in numbers with the growth of annual temperature (Wolfe, 1978).

2. Numerous and diverse cycadophytes in the flora indicate warm growth conditions.

3. Diversity of conifers, especially of advanced groups, can probably be associated with cooling (Samylina, 1974).

4. A predominance of large-leaved plants, platanoids in the first place, testifies to a relatively cold climate (Krasilov, 1979).

5. Plants with *Macclintockia* leaves are typical for relatively thermophilic floras; in the Paleogene this genus is known not only from temperate, but also from subtropical floras of the Poltava type.

6. Warmings permit certain plants, common in the more southern for the Northern Hemisphere regions, to penetrate into northern floras (Meyen, 1984).

7. During climatic warming, a northward shift of boundaries between phytoclimatic zones occurs in the Northern Hemisphere.

The Grebenka taphoflora has exhibited numerous thermophylic plants, such as *Cycadites*, *Nilssonia*, entire-margined *Magnoliaephyllum*, *Araliaephyllum*, "*Lindera*," etc. The Grebenka taphoflora contains sprouts with the scale-like leaves of *Pagiophyllum*, which is a characteristic representative of the Late Jurassic and Early Cretaceous subtropical floras from the European-Sinian Region (Vakhrameev, 1988). Plants of Grebenka flora described as *Schefflerаephyllum* are, in fact, most probably representatives of the *Debeya* genus, which is a widespread element of the Late Cretaceous subtropical phytoclimatic zone (Krasilov, 1979). These facts indicate a climatic warming in the Cenomanian following the thermal minimum close to the Early and Late Cretaceous boundary (Samylina, 1974).

The Penzhina flora is less thermophilic. It does not contain cycadophytes and features few intact-margined angiosperms (only *Myrtophyllum penzhinense* Herman). Burials are dominated by remains of large tooth-margined leaves of platanoids and numerous and diverse conifers. This flora seems to correspond to a cooling in the Turonian.

In the Coniacian Kaivayam flora large dentate-margined leaves of angiosperms also predominate. However, it reveals more thermophilic elements like *Nilssonia*, *Ctenis*, *Magnoliaephyllum*, *Araliaephyllum*, *Ternstroemites* than the Penzhina. This bears witness to a certain climatic warming in the Coniacian as compared with the Turonian (Herman, 1985).

Climatic fluctuations of lesser amplitude for the Coniacian can be outlined from the northwest Kamchatkan floras (Herman, 1985). A comparison of the early Kaivayam subcomplex with the Penzhina complex and late Kaivayam subcomplex shows a higher proportion of entire-margin *Magnoliaephyllum* and a certain decline of platanoids is observed in the early Kaivayam. *Ternstroemites harwoodensis* (Dawson) Bell, typical for the subtropical Nanaimo flora (Campanian, Nanaimo Group) of Vancouver Island (Western Canada), has been recorded here. Other representatives of the *Ternstroemites* genus also compose a part of the subtropical floras of North America. This is suggestive of a certain climatic warming during the early Coniacian. At present, however, it is impossible to establish whether this warming was a purely local phenomenon in northwest Kamchatka, or was more widespread. Finds of cycadophytes in the early Coniacian taphoflora from Pekul'nei Ridge seem to confirm the latter.

The Barykov flora is dominated by *Macclintockia* and "*Quercus*" *tschucotica* with small coriaceous leaves. Large dentate-margined leaves are either not numerous nor entirely lacking. The common composition of this flora comprise cycadophytes and entire-margined *Magnoliaephyllum*, *Araliaephyllum*, including *A. polevoii* (Krysht.) Krassilov, which is a species dominating the Gilyak subtropical flora from Sakhalin (Krasilov, 1979). This flora corresponds to a significant climatic warming which appears to have set in the Santonian (the Valizhgen flora of Kamchatka).

The flora of the Gornaya stage still contains cycadophytes (in the Kakanaut taphoflora from the Pehul'nei Lake area they are rather numerous), but angiosperms are dominated by plants with dentate-margined leaves, and the number of platanoids is increasing in comparison with the Barykov stage flora (*Platanus*, *Arthollia*). Plants with coriaceous *Macclintockia* leaves are absent. This flora might have existed in a colder climate than the Barykov stage flora.

The Koryak stage flora is noted for a still more enhanced role of large-leaved angiosperms (*Platanus*, *Arthollia*, *Pseudoprotophyllum*, *Corylus*). Angiosperms

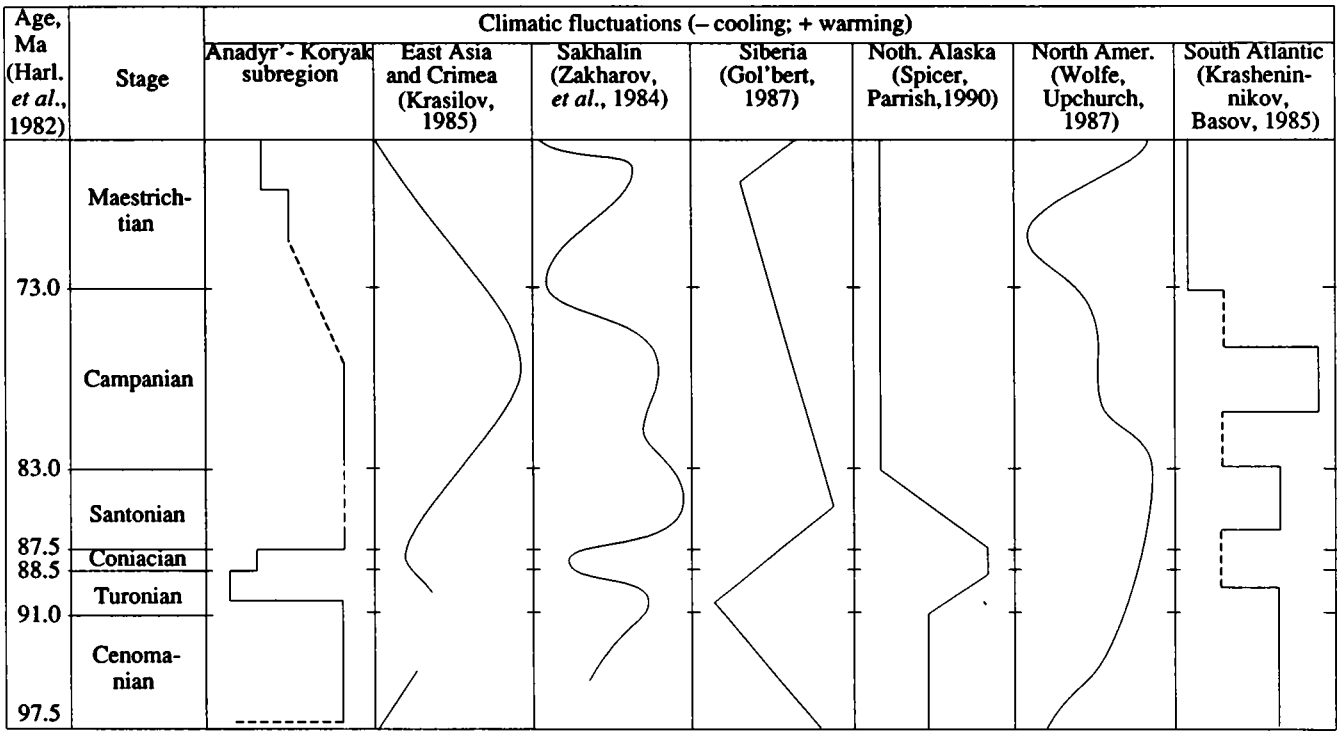


Fig. 5. *Late Cretaceous climate changes (warmings and coolings) for various regions of the Earth (only relative temperature changes are shown).*

with entire-margin leaves are not numerous and cycadophytes are absent. The modern genera of conifers appear to be very important. These features suggest a more substantial climatic cooling during transition from the middle to the late Maestrichtian and, according to the data of L.B. Golovneva (1990) in the Danian.

Thus, the fossil flora of AKSR helps to reconstruct two Late Cretaceous climatic cycles including a warming and a cooling (Fig. 4). The former occurred in the Cenomanian-Coniacian, the latter, in the Santonian-Maestrichtian (Danian). However, as was shown above, taphofloras from the second half of the Campanian and early Maestrichtian are not known on the territory of AKSR (Fig. 2). Therefore, the impression of unidirectional process of cooling after the Santonian to early Campanian optimum till the end of the Cretaceous may prove to be incorrect. It is not impossible that a climatic deterioration occurred in the end of the Campanian and/or early Maestrichtian, while the rather thermophilic middle Maestrichtian flora corresponds to the outset of the third climatic cycle. The data on a relative rise in temperatures in the middle Maestrichtian are present in Kauffman's work (1986).

These climatic fluctuations appear to have been complicated by fluctuations of a higher order. This may be corroborated by a minor warming in the early Coniacian followed by a cooling in the late Coniacian revealed on the basis of fossil flora from northwest Kamchatka.

Climatic cycles were probably associated with cyclic paleofloristic changes reported from the Late Cretaceous of AKSR. Thus, the epochs of warming (Cenomanian, Santonian to early Campanian) are characterized by the presence of cycadophytes, a decline in diversity of conifers and the insignificant role of platanoids. The epochs of cooling (Turonian-Coniacian, late Maestrichtian), are noted for a great diversity of conifers (especially their advanced groups), predominance of large-leaved Platanoids, elimination of cycadophytes, relict representatives of Ginkgoales, Czekanowskiales (during the transition from the Grebenka flora to the Penzhina), and conifers (during transition from the Gornaya flora to the Koryak complex). The coolings must have exerted a positive influence on the *Metasequoia* + *Trochodendroides* "duet." In the Turonian these plants become numerous in taphofloras for the first time, while in the Maestrichtian they come to predominate.

Such a cyclicity of paleofloristic changes, naturally, renders phytostratigraphic correlation more difficult and there is always the danger of confusing floras corresponding to the same parts of different cycles. The long-standing dispute on the age of the Arkagala flora and its place in successive Cretaceous floras of Northeast Asia may be associated with cyclicity. This flora, known from the Arkagala Formation in the upper reaches of the Kolyma River (Verkhoyansk-Chukotka subregion) and from a number of localities within the Okhotsk-Chukotka volcanic belt is characterized (Samylina, 1988) by the predominance of conifers, particularly representatives of the *Pinaceae* and *Taxodiaceae* families. *Sequoia*, *Metasequoia*, *Cryptomeria*,

Cunninghamia, *Cephalotaxus*, *Taxites*, and *Thuja* are numerous. Angiosperms are diverse in systematic composition, but their remains have been found very seldom (except for *Quereuxia angulata* (Newb.) Krysht.) The small-sized leaves of angiosperms are common, platanoids are absent. The Arkagala taphofloras are known for large numbers of fossil *Phoenicopsis* ex gr. *angustifolia* Heer. Among Czekanowskiales, *Czekanowskia* and *Leptostrobus* have been reported. The role of Equisetales, ferns, cycadophites and Ginkgoales is insignificant.

V.A. Samylina (1988) considers the Arkagala flora to be older than the Grebenka, dating it as early Cenomanian. Rarity of angiosperms and predominance of conifers is related to the cold climate in which this flora existed. Other researchers (Krasilov, 1979, Filippova, 1990) assess the Arkagala flora as Turonian, and thus younger than the Grebenka complex. They draw attention to the first appearance in the flora of *Protophyllocladus*, *Metasequoia*, *Trochodendroides*, and *Quereuxia*.

However, if the Arkagala flora is compared to the obviously Turonian Penzhina flora of AKSR, the older appearance of the former becomes evident. This is borne out by low frequencies and small leaves of angiosperms, abundant remains of *Phoenicopsis*, and the presence of a number of relict plants (*Taeniopteris*, *Ginkgo* ex gr. *sibirica*, *Sphenobaiera*, *Pseudotorellia*, *Czekanowskia*, etc.) in the Arkagala flora. It may be assumed to have existed during climatic cooling preceding the Cenomanian optimum. The presence in its composition of advanced genera of conifers and angiosperms, widespread on the territory of AKSR from the Turonian onwards, is connected with climatically dependent cyclic evolution of the Cretaceous floras from northeast Asia and with similar relatively cold conditions of growth near the Albian/Cenomanian boundary, as well as in the Turonian and Coniacian. Edaphically specific isolated depression hosting the Arkagala flora may serve as another explanation of these peculiarities. But in this case the problem hardly can be solved analyzing the floras of only Northeast Russia.

Climatic fluctuations (coolings and warmings) similar to those which affected the development of the Late Cretaceous flora of AKSR are also noted by various researches for other regions of the Earth (Fig. 5). They deserve attention as a possible manifestation of fairly large-scale climatic changes. The following episodes are noteworthy.

1. According to paleobotanic data, a warming in the middle Senomanian is determined for the Crimea (Krasilov, 1985). From the late Albian till early Turonian a warming is observed on the Falkland Plateau (Krasheninnikov, Basov, 1985).

2. A temperature fall occurred in the early Turonian in Siberia (Gol'bert, 1987). The Falkland Plateau is noted for a cooling from the second half of the Turonian till the early Santonian (Krasheninnikov, Basov, 1985). The Coniacian cooling is determined in east Asia (Krasilov, 1985) and in Sakhalin (Zakharov *et al.*, 1984). The Turonian and Coniacian coolings appear to be associated with the expansion of large-leaved Platanoids over the vast territory of the North Pacific (Khatanga River, Vilyui depression, Novosibirsk Islands, Northeast Russia, Sakhalin, Alaska) (Herman, 1991). This expansion could not have been caused only by climatic cooling. It has been suggested (Herman, 1990) that the large size of plant leaves (in certain Turonian platanoids reaching 50 cm) is associated with their vegetation period having proceeded during prolonged, comparatively weak, and largely dissipated sunlight characteristic of high latitudes. The deciduous nature of these plants is probably to be attributed not so much to the cold as to the shortage of light in winter.

3. The warming beginning in the Santonian and continuing at least into the first half of the Campanian appears to have been global or subglobal. It has been re-

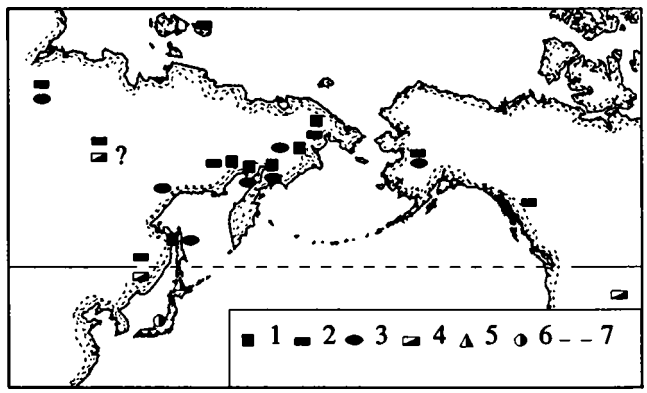


Fig. 6. *Phytogeographic subdivision of the North Pacific region in the Cenomanian-Coniacian;*
1 - 6 – fossil plants localities: 1 – Paraprotophyllum, 2 – Pseudo-protophyllum, 3 – Dalembia, 4 – Protophyllum, 5 – tree ferns, 6 – palms, 7 – southern boundary of the Siberian-Canadian area.

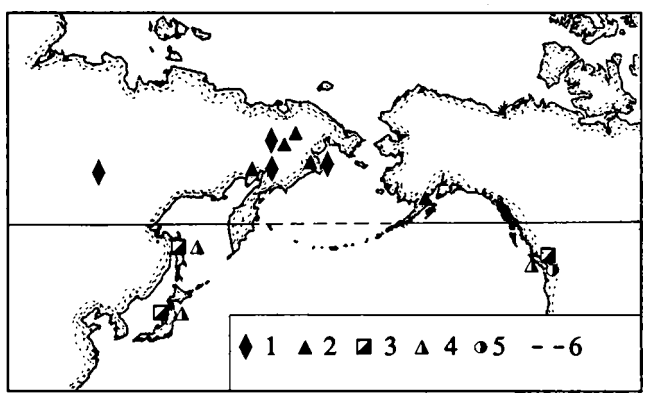


Fig. 7. *Phytogeographic subdivision of the North Pacific region in the Santonian and early Campanian;*
1 - 5–fossil plants localities: 1 – Macclintockia, 2 – "Quercus" tchucotica Abram. ("Q" pseudomarionii Hollick), 3 – Debeya, 4 – Liriophyllum, 5 – palms, 6 – southern boundary of the Siberian-Canadian area.

ported for southern Sakhalin, (Zakharov *et al.*, 1984) with the warmest spell of the Late Cretaceous in this region occurring in the middle Santonian. The Siberian thermal maximum has been recorded in the Santonian (Gol'bert, 1987). A feeble Santonian and a stronger middle Campanian warming has been noted for the region of Falkland Plateau (Krasheninnikov, Basov, 1985). Data from North America testify (Wolfe, Upchurch, 1987) to a gradual warming from the beginning until the middle of the Late Cretaceous, attaining its peak in the Santonian, replaced by a slight cooling in the Campanian. The warming in question appears to have caused a decline in platanoids and the predominance of relatively small- and narrow-leaved angiosperms in the Santonian-Campanian floras from north and northeast Asia, Kazakhstan, Sakhalin, Alaska and Western Canada (Herman, 1991). The warming and certain aridization of climate may be presumed to have reduced cloud cover over North Pacific coastal areas, thus increasing the proportion of direct solar radiation within its overall stream, which must have affected the spreading of large-leaved platanoid plants.

The Santonian-Campanian warming is also obvious from the northward shift of the southern boundary of the Siberian-Canadian paleofloristic region (this boundary is virtually of phytoclimatic nature (Vakhrameev, 1988). In the Santonian and early Campanian this boundary (Fig. 7) moved approximately 1000 km north of its previous position in the Cenomanian-Coniacian (Fig. 6), which corresponds to a warming of about 3°C (Herman, 1991).

4. The cooling in the Maestrichtian, possibly beginning from the late Campanian, is also distinctly traceable on the diagrams (Fig. 5) and appears to have occurred on a large scale. It is to be noted, however, that there is divergence of opinion among various researchers on climatic changes in the second half of the Maestrichtian, i.e. a continued cooling has been reconstructed for east Asia and Falkland Plateau (Krasilov, 1985; Krasheninnikov, Basov, 1985), while a warming toward the end of the Cretaceous has been reported for Siberia and North America (Gol'bert, 1987; Wolfe, Upchurch, 1987). The above evidence on the Late Cretaceous flora of AKSR supports the former view. Maestrichtian warming occurred probably in the middle Maestrichtian and was accompanied by the flora of the Gornaya stage. It was followed by the late Maestrichtian cooling which corresponds to the Koryak stage flora. A similar situation is reconstructed on molluscan fauna of the south Sakhalin (Zakharov *et al.*, 1984).

REFERENCES

Devyatilova, A.D., Nevretdinov, E.B., Filippova, G.G., Stratigraphy of Upper Cretaceous Deposits in the Basin of the Middle Reaches of the Anadyr' River, *Geologiya i Geofizika*, 1980, no. 12, pp. 62 - 70.

Filippova, G.G., Late Cretaceous Floras of Northeastern Asia, *Kontinentalnyi Mel SSSR*, DVO AN SSSR, Vladivostok, 1990, pp. 127 - 131.

Gol'bert, A.B., *Osnovy Regional'noi Paleoclimatologii*, (Fundamentals of Regional Paleoclimatology), Nedra, Moscow, 1987.

Golovneva, L.B., Maestrichtian-Danian Flora of the Rarytkin Ridge (Koryak Highland), Cand. Sci. (Biol.) Dissertation, Botanical Institute, USSR Academy of Sciences, Leningrad, 1990.

Harland, W.B., Cox, A.V., Llewellyn, P.G., *et al.*, *A Geologic Time Scale*, Cambridge Univ., 1982.

Herman, A.B., A First Find of the *Ternstroemites* Genus in the USSR (Late Cretaceous of Kamchatka), *Paleontol. Zhurnal*, 1985, no. 1, pp. 138 - 141.

Herman, A.B., Cenomanian Flora of the Ugol'naya Bay (Northeast of the USSR), *Izvestiya AN SSSR, Ser. Geol.*, 1988, no. 11, pp. 110 - 114.

Herman, A.B., Late Cretaceous Floras and Climate of the Anadyr'-Koryakian Subregion (North-East USSR), *Proc. Symp. Paleofloristic and Paleoclimatic Changes in the Cretaceous and Tertiary*, Prague, 1989; *Geol. Surv.*, Prague, 1990b, pp. 73 - 79.

Herman, A.B., Paleobotanic Substantiation of Correlation of Upper Cretaceous Deposits in the Northeast of the USSR, *Yarusnye i Zonalnye Shkaly Boreal'nogo Mezozoya SSSR*, Nauka, Moscow, 1989, pp. 204 - 212.

Herman, A.B., Changeability of Leaves and Recon- struction of an Annual Sprout of *Paraprotophyllum ignatianum*, *Regional'naya Geologiya: Tektonika, Litologiya i Stratigrafiya*, GIN AN SSSR, Moscow, 1990, pp. 79 - 83.

Herman, A.B., Cretaceous Angiosperms and Phytostratigraphy of Northwest Kamchatka and Elistratov Peninsula, *Stratigrafiya i Flora Melovych Otlozhenii Severo-Zapadnoy Kamchatki*, Nauka, Moscow, 1991, pp. 5 - 141.

Kauffman, E.J., Structure of Extinction of Marine Biotas in the Cretaceous, *Katastrofy i Istoriya Zemli, Novyi Uniformizm*, Mir, Moscow, 1986, pp. 156 - 254.

Krasheninnikov, V.A., Basov, I.A., *Stratigrafiya Mela Yuzhnogo Okeana*, (Cretaceous Stratigraphy of the Southern Ocean), Nauka, Moscow, 1985.

Krasilov, V.A., Evolution of Late Cretaceous Vegetation on the Western Pacific Coast in Relation to Climatic Changes and Tectogenesis, *Iskopaemye Flory Dal'nego Vostoka*, DVNTS AN SSSR, Vladivostok, 1976, pp. 30 - 42.

Krasilov, V.A., *Melovaya Flora Sakhalina* (Cretaceous Flora of Sakhalin), Nauka, Moscow, 1979.

Krasilov, V.A., *Melovoi Period. Evolutsiya Zemnoi Kory i Biosfery* (Cretaceous Period. Evolution of the Earth's Crust and Biosphere), Nauka, Moscow, 1985.

Meyen, S.V., Phylogeny of Higher Plants and Florogenesis, *Doklady 27-go Mezhdunar. Geol. Kongressa*, Nauka, Moscow, 1984, vol. 2, Paleontologiya, pp. 75 - 80.

Nesov, L.A., Golovneva, L.B., History of the Evolution of flora, Vertebrate Fauna, and Climate in Late Senonian in the Northeastern Koryak Highland, *Kontinentalnyi mel SSSR*, DVO AN SSSR, Vladivostok, 1990, pp. 191 - 212.

Samylina, V.A., Early Cretaceous Floras of the Northeastern USSR, (On the Problem of the Formation of Cenophytic Floras), *XXVII Komarovskie Chteniya (16 Oktyabrya 1973)*, 27th Komarov Readings (October 16, 1973), Nauka, Leningrad, 1974.

Samylina, V.A., *Arkagalinskaya Stratoflora Severo-Vostoka Azii*, (Arkagala Stratoflora of Northeastern Asia), Nauka, Leningrad, 1988.

Shchepetov, S.V., Herman, A.B., Cretaceous Flora of the Anadyr' River Right Banks, *Izvestiya AN SSSR, Ser. Geol.*, 1990, no. 10, pp. 16 - 24.

Shchepetov, S.V., Herman, A.B., Belaya, B.V., *Srednemelovaya Flora Pravoberezh'ya Reki Anadyr' (Sistematicheskii Sostav, Stratigraficheskoe Polozhenie, Atlas Iskopaemoi Flory)*, (Middle Cretaceous Flora of the Anadyr' river Right Banks (Systematic Composition, Stratigraphic Position, Atlas of Fossil Flora)) SVKNII DVO AN SSSR, Magadan, 1992.

Spicer, R.A., Parrish, J.T., Late Cretaceous – Early Tertiary Paleoclimates of Northern High Latitudes: a Quantitative View, *Journ. of Geol. Soc.* London, 1990, vol. 147, no. 2, pp. 329 - 341.

Terekhova, G.P., Filippova, G.G., On the Structure and Age of the Poperechnaya Formation (Pekul'nei Ridge, Northeastern USSR), *Dokl. AN SSSR*, 1983, vol. 269, no. 4, pp. 911 - 914.

Terekhova, G.P., Filippova, G.G., Tylpegyrgynai Formation and Its Floral Complex (Pekul'nei Ridge, Northeastern USSR), *Doklady AN SSSR*, 1984, vol. 278, no. 6, pp. 1443 - 1446.

Vakhrameev, V.A., *Yurskie i Rannemelovye Flory Evrazii i Paleofloristicheskie Provintsii Etogo Vremeni* (Jurassic and Early Cretaceous Floras of Eurasia and Paleofloristic Provinces of This Time), Nauka, Moscow, 1964.

Vakhrameev, V.A., *Yurskie i Melovye Flory i Klimaty Zemli* (Jurassic and Cretaceous Floras and Climates of the Earth), Nauka, Moscow, 1988.

Wolfe, J.A., A Paleobotanical Interpretation of Tertiary Climates in the Northern Hemisphere, *Amer. Sci.*, 1978, vol. 6, no. 66, pp. 693 - 703.

Wolfe J.A., Late Cretaceous - Cenozoic History of Deciduousness and the Terminal Cretaceous Event, *Paleobiology*, 1987, vol. 13, no. 2, pp. 215 - 226.

Wolfe, J.A., Upchurch, G.R., North American Nonmarine Climates and Vegetation During the Late Cretaceous. *Palaeogeogr., Palaeclimatol., Palaeecol.*, 1987, vol. 61, no. 1 - 2, pp. 33 - 77.

Zakharov, Y.D., Grabovskaya, V.S., Kalishevich, T.G., Late Cretaceous Succession of Marine Communities in southern Sakhalin and Climatic Features of the Northwest Pacific, *Sistematika i Evolyutsiya Bespozvonochnykh Dal'nego Vostoka*, DVNTS AN SSSR, Vladivostok, 1984, pp. 41 - 90.

Stratigraphy and Geological Correlation, Vol. 1, No. 1, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 1, 1993, pp. 97 - 108.
Original Russian Edition Copyright © 1993 by Stratigrafiya. Geologicheskaya Korrelyatsiya, Gladenkov, Shantser.
English Translation Copyright © 1993 by Interperiodica.

Evolution of the biosphere

Paleogene Geological Events in Kamchatka

Yu. B. Gladenkov and A. E. Shantser

Geological Institute, Russian Academy of Sciences, Pyzhevskii Per. 7, Moscow, 109017, Russia

Manuscript received April 15, 1992

Abstract – Various geological events that occurred in Kamchatka in Paleogene time were correlated for the first time using a new stratigraphic scheme: faunal and floral changes, climatic fluctuations, tectonic movements, volcanic outbursts, sea transgressions, and regressions. Turning points were established in the geological evolution of the region. Stages in the occurrence of geological events and their relative synchroneity were outlined.

INTRODUCTION

One of the fundamental objectives of geology is to define the synchroneity and interrelation of various geological events, both on regional and subglobal scales. Unfortunately, the insufficiently investigated regions make it difficult to study the problem completely. Therefore, the synchronization of transgressions and regressions, tectogenesis phases, strong outbursts of volcanoes, and other geological phenomena is still provisional. This is true of the northern Pacific Ocean surroundings, and this paper is devoted to one of its segments.

The combined investigations (stratigraphic, lithologic, and tectonic) that were recently carried out by the Geological Institute of the Russian Academy of Sciences in Kamchatka have thrown new light on many aspects of the evolution of the region.

Several years ago, we analyzed and compared geological events of one of the most thoroughly studied Cenozoic intervals, the Neogene (Gladenkov and Shantser, 1989). Major sedimentation cycles, volcanic and tectonic events, paleobiota changes, climatic fluctuations, etc. were correlated using a single stratigraphic background based on recent data. Such a calendar of geological phenomena, prepared in the form of a stratigraphic scheme of events, made it possible to reveal some synchroneity between them and, in several cases, even to determine cause and effect relations. It was more difficult to decipher events of the Paleogene, which is less studied than the Neogene. Until recently, the idea of continuous sedimentation in the Paleogene of Western and Eastern Kamchatka dominated in the literature without being confirmed by corresponding evidence; the concept treated the volcanism here, which was *a priori* regarded to be of the island-arc nature, in a somewhat simplified way.

The data obtained during the last 10 to 15 years allow us to correct our previous conceptions. To begin with, new versions of stratigraphic schemes have been drawn for the Paleogene of many regions of Kamchatka with due regard for the materials on the restudied and new-opened reference sections. In Western Kamchatka, these are the Chemurnaut, Palana, Tochilo, Mainach, Utkholok, and Snatol sections (sections are named after local geographic localities); in Eastern Kamchatka, they are the Il'pinskii, Cape Kamchatka, Kronotskii, and some other sections. It is to be noted that the study of these sections was (in many cases) accompanied by monographic description of fauna and flora; this enabled us to define the age of Paleogene horizons more precisely than it had been done previously (Gladenkov *et al.*, 1985; 1991; Gladenkov *et al.*, 1988; Gladenkov *et al.*, 1990; Chelebaeva and Brattseva, 1985). During the same period, an important step was made in the investigation of planktonic remnants (foraminifers and partially nannoplankton and diatoms), enabling us to divide the Paleogene sequences into zones and, on this basis, to carry out broad correlations. The most significant and complete section appeared to be Il'pinskii; seven zones were identified in it in the Paleocene to upper Eocene. The close connections between lonas and beds with fauna identified according to benthic assemblages and these zones allow us to apply the zonal scheme to many regions. Now these zones have also been identified in the Cape Kamchatka and Kronotskii peninsulas. The study of diatomic marine complexes (Gladenkov, 1991) enabled us to describe the Oligocene, which had been always difficult to identify. In particular, it was exposed more completely in Karaginskii and the Bering Islands. All the cited data considerably changed the dating of the sequences in Kamchatka and, therefore, compelled us

to review our former conceptions about the continuity of the sections in the region. On the basis of direct geological observations (searching for angular unconformities) and biostratigraphic data, we determined that the lower - early - middle Eocene hiatus is the largest one both in Western and Eastern Kamchatka. This interval accounts for the high surface level, the intense tectonic deformations, and the topographic planation.

The analysis of Paleogene sequences in Kamchatka revealed long-term sedimentation cycles that are generally synchronous with the epochs of continental and submarine volcanism (volcano-sedimentary cycles – Shantser, 1979). At the same time, the different character of the sedimentation and volcanism in various parts of Kamchatka suggests that the Western Kamchatka subregion (together with central Kamchatka) belongs to the area of unstable orogenic development, whereas Eastern Kamchatka is generally close to island-arc and suboceanic regions.

Figure 1 shows the major Paleogene events in Kamchatka on the basis of a single stratigraphic scheme of new generation. This specific calendar of events does not only record their succession; it also enables us to outline focal points in geological history, which reflect the synchroneity of various phenomena. Owing to the specific nature of the Western and Eastern Kamchatkan geology, we should analyze Paleogene events in these subregions separately.

THE WESTERN KAMCHATKA SUBREGION

Two large stages of development are clearly distinguished in the Paleogene: the Paleocene to early Eocene and the middle Eocene to Oligocene. They are separated by an interruption in sedimentation, which coincides with the late early Eocene and early middle Eocene time.

The Paleogene - Early Eocene Stage (The 1st Volcano-Sedimentary Cycle)

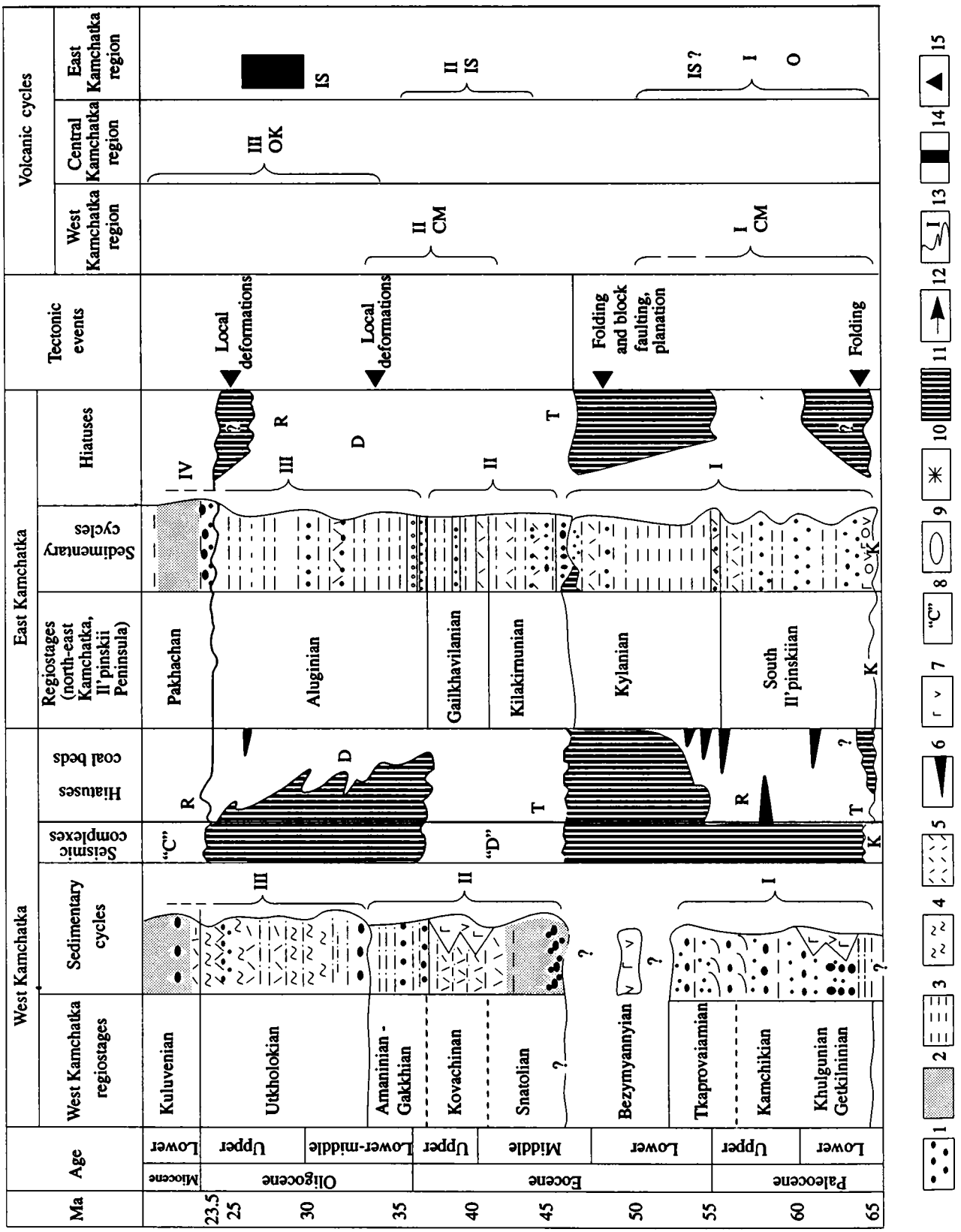
This stage is marked by a characteristic series of shallow-water coastal and continental marine deposits. The deposits of this stage have been most completely studied in the Tigil' area and in the northwest of the region, along the eastern coast of Penzhina Bay.

In the first case, these are terrigenous and volcaniclastic sediments, which are often carbon-bearing, with a large amount of silty material, and locally distributed beds of coarse psephites. The sections studied display alternate salt- and fresh-water mollusc remnants. Mass burial of leaf flora is a characteristic feature here. In the Utkholok Peninsula, vertical and lateral conformable transitions can be observed between the terrigenous formations and the sequence of subaerial, and locally subaqual volcanic rocks of the calc-alkalic series, dominated by andesites and basaltic andesites. There is a great amount of pyroclastics: agglomeratic, psephitic, and psammitic tuffs, lahars, and volcanogenic proluvium.

The stratigraphy of Paleocene strata in the Tigil' area is based on paleoflora analysis (Chelebaeva and Brattseva, 1985). Two phytohorizons have been identified in the Ulevenei Group: the Khulgun (supposed to be of Danian age) and the Sosopkhan (late Paleocene). They are dated by comparing their flora with that of the adjacent regions of Alaska and Sakhalin. It is to be noted that Paleocene beds are poorly exposed in this region (except for the Utkholokian Peninsula coast and, to a lesser extent, in the Snatol and Kovachina river valleys). Sections must often be combined from scattered outcroppings in numerous tectonic blocks. If it were not for rich flora burials, it would be difficult to compile the section. We did not manage to discover the lower contact of the sequence with the underlying Cretaceous formations. As for the upper contact with the Snatol and Kovachina middle Eocene formations, it was determined only in Kvachina Bay near Cape Ovra, where an angular unconformity is clearly observed. It is not easy to determine the real thickness of the lower Paleogene sequences in the region. There are blocks in which the thickness of the deposits containing Sosopkhan flora is estimated to be as much as 1000 to 1200 m. In the Utkholok Peninsula section, the apparent thickness of the terrigenous formations (with the Khulgun floral complex) is 1000 m. Therefore, we may conclude that the sedimentary strata accumulated in the Tigil' area during the Paleocene is more than 2000 m thick.

The area, over which Paleocene formations are spread, consists of a complicated system of tectonic blocks, inside which the strata form monoclines with angles of dip reaching, on average, 20 to 30° and up to 60 to 70° in narrow nearfault zones: all the dips are of the eastern vergence. More intense deformations are observed along the western coast of the Utkholok Peninsula. Subvertical occurrences are frequent here. There are lots of upthrow faults, and a hinge overthrust with an amplitude of allochthon displacement up to 4 to 5 km was recorded here. Subaerial volcanic rocks are thrust over terrigenous strata (Danian?) in an eastward direction.

On the eastern coast of Penzhina Bay, deposits, which are referred to the early Paleogene stage, are clearly exposed in coastal escarpments along a distance of tens of kilometers. They are mostly represented by terrigeneous formations with a large amount of tuffaceous material. In the continuous section (the Getkilnin, Kamchik, Tkapravayam formations of the Chemurnaut Group), we observe a gradual transition from marine deposits through coastal and subcontinental ones to purely continental deposits. Accordingly, the amount of coarse material increases upsection. Several dozen normal alluvial cycles, with a large conglomerate component, have been identified in the upper part of the section. Its lower part is dominated by sandy-clayey facies. The beds are characterized by low-grade coal occurrences. Deposits that accumulated under avalanche sedimentation are typical of the middle and upper parts of the section; it is characterized in particular by petrified and charred sunk logs penetrating


Ma
Age
West Kamchatka
West Kamchatka regiostages
Sedimentary cycles
Seismic complexes
Hiatuses coal beds
East Kamchatka
Regiostages (north-east Kamchatka, Il'pinskii Peninsula)
Sedimentary cycles
Hiatuses
Tectonic events
Volcanic cycles
West Kamchatka region
Central Kamchatka region
East Kamchatka region
Miocene
Oligocene
Eocene
Paleocene
Lower
Upper
Lower-middle
Middle
Kuluvenian
Utkholokian
Amaninian - Gakhkian
Kovachinan
Snatolian
Bezymyannyian
Tkaprovaiamian
Kamchikian
Khulgunian
Gekilinian
Pakhachan
Aluginian
Gailkhavilanian
Kilakirnunian
Kylanian
South Il'pinskiian
Local deformations
Folding and block faulting, planation
Folding
OK
CM
IS
O

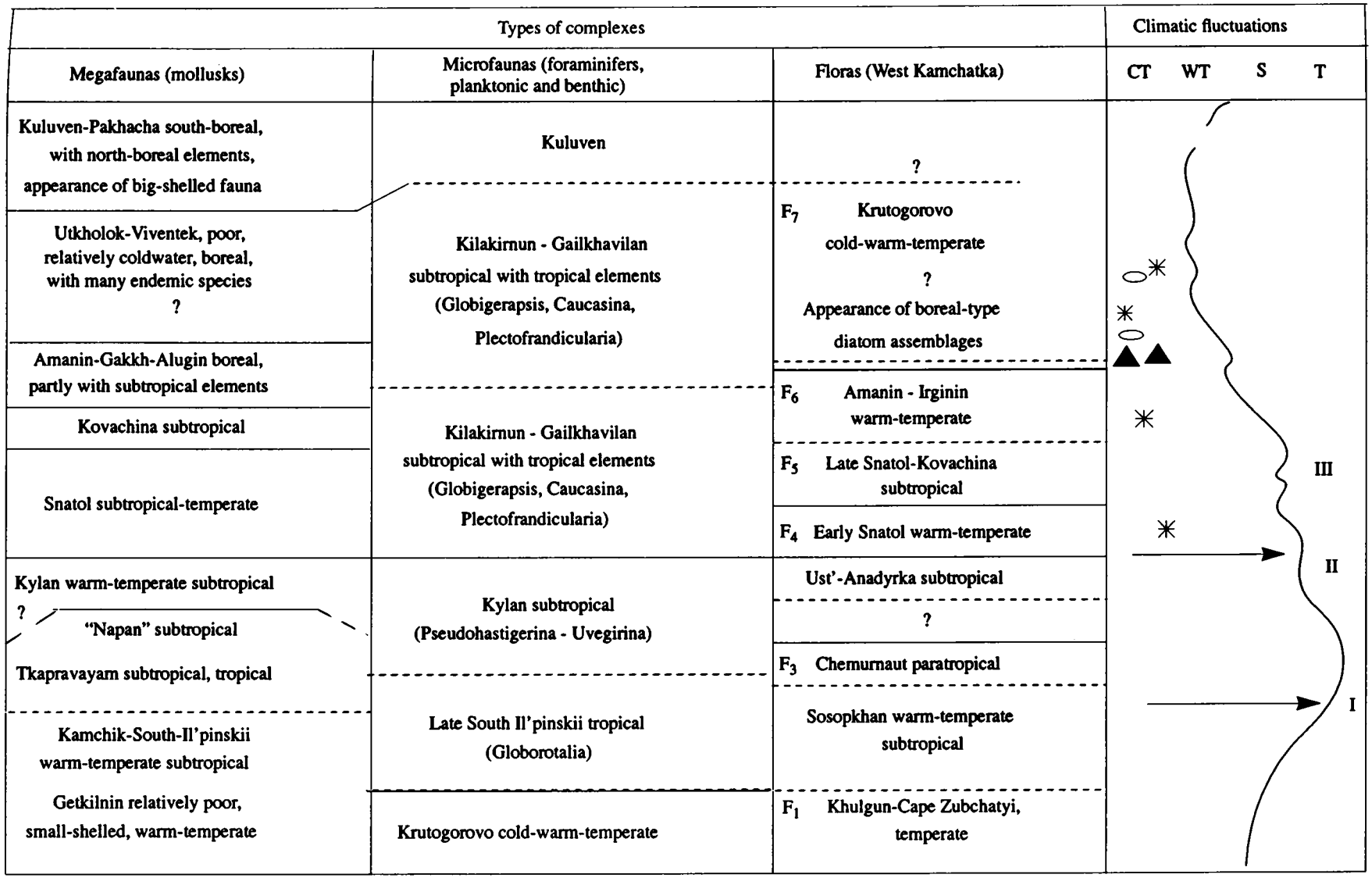


Fig. 1. *Correlation of Paleogene Geological Events in Kamchatka.*
1 – Conglomerates, 2 – sandstones, 3 – siltstones and claystones, 4 – siliceous rocks, 5 – interbeds of tuffs in the sedimentary sequence; 6 – coals; 7 – basalts, andesites, and basaltic andesites, 8 – seismic complexes of the Sea of Okhotsk, 9 – concentrations of floating pebbles and boulders, 10 – gennoishes, 11 – stratigraphic hiatuses, 12 – molluscan fauna migration, 13 – climatic curve, 14 – the Aleutian island arc formation; 15 (D) – mass appearance of diatoms, T – sea transgressions, R – sea regressions, CM – continental margin volcanic rocks, IS – island arc volcanic rocks, O – oceanic volcanic rocks, climate: CT – cold-temperature, WT – warm-temperate, S – subtropical, T – tropical.

through a series of beds. There are strata of lateral accretion with coarse lenticular cross bedding. The total thickness of the group is estimated to be as much as 4000 m.

The Chemurnaut Group sequence is abundant in sills and cutting dikes. The composition of the sills varies from basalts to dacites. Sills and dikes evidently served as a feeder system for the later (Eocene) zone of terrestrial volcanism, which is confirmed by absolute datings. The Chemurnaut Group may be interpreted as a single regressive sedimentation cycle. But the nature of the group's lower contact with Cretaceous formations, such as that in the Tigil' area, remains unclear. The upper contact is exposed in the Tanuinginan Bay, where the Tkapravayam Formation deposits are overlain with a sharp angular unconformity by volcanic sedimentary facies of the Bozhedom Formation, which (according to paleoflora analysis) may be attributed either to the upper Middle Eocene or to the upper Eocene.

Until recently, the age of the Chemurnaut Group was defined by paleoflora. L.Yu. Budantsev (1984) identified a floral assemblage in the group, which he assigned to the Eocene. Later, a comparative analysis of the Alaskan and Chemurnaut floras enabled A.I. Chelebaeva (Chelebaeva, Brattseva, 1985) to refer them to Paleocene or early Eocene, assuming that the section can be built up (i.e., by identifying conventionally the marine Getkilnin Formation with the lower part of the Ulevenei Group and the Kamchin Formation with the upper Ulevenei). As to the Tkapravayam flora, which is similar to the paratropic type, Chelebaeva correlates it with the lower Eocene climatic optimum. The data on marine benthic assemblages generally corresponds well to that cited above. M.Ya. Serova (Serova *et al.*, 1989) recorded remains of late Danian foraminifers in the Getkilnin Formation. According to V.N. Sinel'nikova and Yu.B. Gladenkov, the mollusks from the Chemurnaut Group and from the Ulevenei beds, if compared with those from assemblages in other regions, are likely to belong to the Lower Paleogene (Paleocene).

The rocks of the Chemurnaut Group form a simple extended synclinal fold, which extends northeastward with gently dipping rocks in its core. Its western limb contains numerous faults, and, in many parts, dips steeply up the vertical occurrence. A series of faults, believed to be normal faults, can be clearly traced along the Penzhina Bay coast in a northeasterly direction. The above-mentioned angular unconfurmity leads us to conclude that the original deformations are of middle Eocene age.

The reported data indicate that the beds of the first early Paleogene period of the Western Kamchatka evolution may be generally assigned to the Paleocene and the early Eocene. Now, it is logical to state that there existed a shallow-water basin in the Paleocene, which periodically dried up. It is likely that it also occupied the small areas of the western part of the Okhotsk aquatorium because the ocean drilling data (Shainyan, 1987), indicate that the Cretaceous deposits are directly overlain by middle-upper Eocene strata. Evidently, active subaerial and subaqueous calc-alkalic volcanism took place within the sedimentary basin and, possibly, along its surroundings.

Paleocene volcanic sequences, recorded in the Utkholok Peninsula, come into contact with terrigeneous flora-bearing formations. Volcanic rocks are cut off by the present coast of the Gulf of Shelekhov; therefore, we may suggest the existence of a volcanic zone in the Paleocene within the gulf's aquatorium and its subsequent submergence. However, there are data on the Paleocene volcanism in the mainland part of the Kamchatka Peninsula, in the basins of the Shamanka and Lesnaya rivers. The molasse known in this region occurs on a planated surface, developed on terrestrial volcanic rocks and comagmatic granitoid intrusions. Fauna and flora analysis showed that the molasse is of Middle-Late Eocene age (Shantser, 1987). Furthermore, our data indicate that the formation of terrestrial volcanic rocks occurred simultaneously with the deposition of the clayey strata on the marine shoal on the coast of the Gulf of Shelekhov at the same latitude. By lithological features and distribution areas, these strata are commonly identified in geological surveys with the Getkilnin Formation of the Chernaut Group (Late Danian?). Apart from datings of volcanic sequences, whether they are correct or not, a considerable amount of juvenile material and redeposited tephra indicates intense and long-term volcanism in the Paleocene in the western part of the region, which is apparently a decisive factor for intense sedimentation in the shallow-water basin and is apparently characterized by an oscillating regime of subsidence. Growing stratovolcanoes, known to be abundant in pyroclastic, under conditions of humid subtropical climate, were rapidly destroyed and eroded, supplying a large amount of clastic material (probably by the turbidite flows) to the neighboring waters. The avalanche sedimentation characteristic of the Chemurnaut Group may be cited as an example. On the whole, the Paleocene in the western part of the region is represented largely by a major volcano-sedimentary cycle with sedimentation similar to the compensated type, which was accompanied by the formation of coals in shallow-water marine and freshwater basins against the background of their discrete subsidence.

A considerable share of endemic molluscs and the absence of radiolarians and calcareous plankton (in spite of the warm subtropical early Paleocene climate) indicate the absence of a stable connection between the Western Kamchatka basin and the eastern oceanic aquatorium and may mark a certain isolation of the former.

Specific Features of the Paleocene-Early Eocene Tectonics in the Western Kamchatka Subregion

The problem of the relationship between Paleocene and Cretaceous formations in western Kamchatka

remains open. Only the Anadyrka River mouth, where the provisionally Paleocene Anadyrka Formation overlies the Upper Cretaceous volcanic deposits (Gladenkov *et al.*, 1990). The opposite vergence observed here, in folds of the Cretaceous and Paleocene beds (in Cretaceous – western, in Paleocene – eastern), leads us to suggest a structural reconstruction at this boundary.

As was mentioned previously, the lower and (to a lesser extent) middle Eocene in the Western Kamchatka subregion were marked by a major hiatus, which is emphasized by the angular unconformity at the base of the Snatol Formation, less frequently of the Kovachina Formation and of their analogues. Apparently this hiatus coincided with a high surface level, intense tectogenesis, and subsequent topographic planation. Among the major tectonic features of this time, the following seems to be most significant. A quite narrow strip along the western coast was characterized by the most intense deformations: frequent vertical occurences of rocks, overturned folds, and short hinge overthrusts. This phenomenon is probably connected with the appearance of an extended fault zone extending northeastward along the Okhotsk coast of Kamchatka and having a differentiated downthrow-upthrow character. It is to be noted that during folding and overthrusting, the eastern vergence dominated: whereas, the western vergence prevailed in Cretaceous time. Finally, the Eocene was marked by an active development of a diagonal northwest fault system, particularly in the Kamchatka Isthmus area (Ozernaya-Khailyuli system, according to A.E. Shantser (1979)). Further to the southeast, it comes into contact with faults that are conjugated with the trend of the Aleutian island arc, which formed a little later. Large-amplitude right-lateral faults presumably occurred along the diagonal fault zone.

The Eocene-Early Miocene Stage.

The next evolutionary stage of the Western Kamchatka subregion comprises the interval from the middle Eocene to the early Miocene, in the sequence, from the Snatol Formation to the Kuluven Formation (Gladenkov and Shantser, 1989; Gladenkov *et al.*, 1991). Previously, we included the deposits of this stage (Shantser, 1987) in a single sedimentation cycle, the middle Eocene to Early Miocene and identified two subcycles in it: the middle Eocene to early Oligocene and the Oligocene to early Miocene. The former includes the Snatol (up to several hundred meters), Kovachina (600 m), Amanina (200 m), and the Gakkh (about 1000 m) formations of Western Kamchatka. The latter includes the Utkholok (450 m), Viventek (300 m), and the Kuluven (about 300 m) formations. Since the latest data suggest a major, though not ubiquitous, hiatus in pre-Utkholok (Oligocene) time, and a simultaneous substantial rearrangement of volcanic zones, it would be more logical to regard the identified subcycles as independent cycles: the second and the third cycles of the Paleogene sedimentation (Fig. 1).

The Second Volcano-Sedimentary Cycle.

This Paleogene cycle began with shallow-sea transgression in Snatolian time; it involved extensive areas of the region. The distribution of mollusks enables us to conclude that, beginning with this time (the second half of the middle Eocene), the connection between the eastern and western sedimentation basins was established, probably through straits in the land that divided the basins. The middle Eocene (the Snatol Formation) is characterized by an accumulation of shallow-sea terrigenous sediments, which have practically no admixture of volcanic material. Therefore, we may suggest a pause in the development of volcanism; it may have started in the period of a high surface level and folding.

Transgression spreading and sea deepening continued in Kovachinan time (the latest Middle-Late Eocene) with a subsequent decrease of depth and a diminishing of the area of the sedimentary basins in pre-Utkholokian time (Oligocene). Considerable silica (mainly organic) precipitation, accompanied by the formation of opokas and opoka-like rocks, took place in Kovachinan and Amaninian-Gakkhian time against the background of essentially terrigenous sedimentation. Simultaneously to marine sedimentation, active terrestrial volcanism occurred again in late Snatolian- Kovachinan time, being inherited (only in the Eocene) from the Paleocene volcanic zone. It is to be noted that reliably dated volcanic rocks of that time are recorded only at the extreme northwest of the Western Kamchatka subregion and on western slopes of the southern part of the Sredinnyi Ridge (Chernaya Mt.). An abundance of tuffs (ashes) and mixed rocks in the Kovachina, Amanin, and Gakkh formations, often making up marker-interbeds, indicates a wider distribution of terrestrial volcanic rocks, possibly in present aquatorium of the Sea of Okhotsk and in the Kamchatka Sredinnyi Ridge area.

Apparently, the influence of volcanism on sedimentation is most clearly pronounced in the Amanin and Gakkh formations (the end of the late Eocene or beginning of the Oligocene). Frequent alternation of small pseudocyclothems (average thickness 0.3 to 0.7 m) can be observed here, with their upper part mostly represented by tuff (frequently desintegrating to become clay), generally of psammitic and pelitic granulometry. Each elementary cyclothem in the sediments may be regarded as corresponding to the cycles of volcanic development, which yield large amounts of pyroclastic material settling down in the littoral and sublittoral area. On the whole, the second cycle of Paleogene sedimentation (just like the Paleocene one), may be called volcano-sedimentary.

The pre-Utcholok hiatus (Oligocene) was not accompanied by angular unconformities. It is indicated by the absence of the Amanin-Gakkh assemblages in the sequence (particularly in the southern areas of Western Kamchatka) and by the partial occurrence of the Utcholok Formation on older deposits (the geolog-

ical survey data), as well as by a sharp coarsening of clastic material within the formation. According to the evidence obtained from a borehole in the northern part of the Sea of Okhotsk, a large hiatus, covering virtually the entire Oligocene, indicates a marked decrease in the area of the aquatorium at that time (Shainyan, 1987).

Apparently the pre-Utkholok hiatus coincided in time with an intense rearrangement of volcanic zones (a "jumping" of the active terrestrial volcanism to the central part of the region) with the beginning of the long-term formation of the Oligocene to middle Miocene Sredinnyi volcanic zone, in which volcanism continued in the Pliocene-Quaternary time after an interruption in the late Miocene.

The Third Volcano-Sedimentary Cycle.

In the nature of sedimentation, the third Paleogene cycle was similar to the second. Rocks also contained a considerable amount of tuffs, composing in many places the upper parts of cyclothems (particularly in the Viventek Formation): the accumulation of substantially siliceous sediments (opokas and diatomites, in the Viventek Formation) continued. Coarsening of sediments and the sea regression did not start until Kuluvenian (early Miocene) time. As volcanism continued to be one of decisive factors of sediment accumulation, the third cycle may be regarded as volcano-sedimentary. Subsynchroneity of the sedimentary and volcanic cycles is shown in the attached scheme (Fig. 1).

Main deformation of the rocks of the second and third Paleogene cycles took place in middle and late Miocene time; it was of synsedimentary (Middle Miocene) and post-sedimentary (Late Miocene) character. More detailed information on the Miocene tectogenesis was given in earlier works (Shantser, 1979, 1987). We would note only characteristic intrastratal dislocations, usually of small amplitude, which generally occurred in ash beds, which were, as a result, transformed into tectonic clays.

THE EASTERN KAMCHATKA SUBREGION

Geologically, this region is much more complicated than Western Kamchatka. It comprises several complex tectonic structures, the deciphering of which is far from complete. Below we attempt to consider the most general aspects of the evolution of this subregion.

Recently, a very important step was made in the study of stratigraphy of Eastern Kamchatka. As was already mentioned, the zonal subdivision of Paleogene sequences was done for the first time here on the basis of planktonic foraminifers and partially on the basis of nannoplankton and diatoms. They characterize the South Il'pinskii, Kylan, Kilakirnun, and Gailkhavilan Formations of the Il'pinskii Peninsula of northeastern Kamchatka. Many of the zones have been recently identified in Cape Kamchatka and the Kronotskii Peninsula.

Though a number of Kamchatka sequences are not yet sufficiently dated, in the Vostochnyi Range, the Karaginskii I. (eastern part), Goven Peninsula, and other areas, a stratigraphic scheme has recently been compiled, which enables us to objectively define the age of past events. Similarly to Western Kamchatka, two large groups of sediments have been identified here, which seem to be divided by an early to middle Eocene hiatus that was recorded in the Kronotskii Peninsula (at the boundary between the Kubov and Kozlov Formations), in the Ozernoi Peninsula, and, less distinctly, in Cape Kamchatka. Middle Eocene rocks, unconformably overlying more ancient sequences, have been observed in the Valaginskii Range.

The Paleogene stratigraphy of Eastern Kamchatka is best illustrated by the most representative section in the Il'pinskii Peninsula. The lowermost parts of the section, the lower part of the South Il'pinskii Formation, may be attributed to the first half of the Paleocene (the radiometric age of the lavas is 54 to 58 Ma; the microfauna is evidently Danian). Upsection (up to 400 m), the rocks are overlain without any visible hiatus by Eocene strata (the Kylan Formation (500 m thick) and other formations). However, paleontological evidence suggests a hidden hiatus at the base of the Kilakirnun Formation, extending through uppermost lower Eocene and lower parts of the middle Eocene.

The lower part of the South Il'pinskii Formation is characterized by terrigenous sediments interbedded with pyroclastics and lavas of basalt composition. Volcanic rocks are also present in the conformably overlying upper subformation. The volcanogenic component increases to the east, and all volcanic rocks here are of submarine genesis. Marine volcanic formations are distributed east of the Il'pinskii Peninsula in the Goven Peninsula; because of the sparse mollusk fauna, they have been provisionally correlated with the lower and upper parts of the Il'pinskii section. Though the correlations are still, in many cases, approximate, we can already speak of the existence of a marine basin in the Paleocene to Early Eocene, with substantially terrigenous sedimentation in the western parts of the peninsula, and predominantly volcanogenic sedimentation in the eastern parts of the peninsula. The volcanic and terrigenous sedimentation zones are conjugated more closely here (developing under similar conditions), than in the western part of the Kamchatka region. A peculiar volcano-sedimentary cycle can be outlined here, which was synchronous with the first Paleogene cycle of Western Kamchatka. In the Eastern Kamchatka areas lying farther south (the Cape Kamchatka, the Kronotskii Peninsula, the Vostochnyi Range, and others) the Paleocene level is associated with submarine volcanism, which is supposed to have developed under shallow-water conditions. Its rocks are low-potassium, high-alumina tholeiites, like (for example) those of the Kubov Formation, which are close to the tholeiites of primitive island arcs (Khubunaya, 1984). We might suggest the existence of a submarine zone of volcanism in the Paleocene to Early Eocene: it extended from the

Goven Peninsula in the north to the Kronotskii and, possibly, to the Shipunskii peninsulas in the south. However, the age and the lithofacies character of the eastern Karaginskii sequences and of the Vetlov Formation in the Vostochnyi Range remain unclear. Isolated finds of benthic foraminifers and radiolarians allow us to define their age from the Maestrichtian to the Eocene. According to the data reported by various authors, these formations contain both primitive basalts, similar to those of mid-oceanic ridges, and alkalic basalts, often melanocratic, as well as acidic volcanic rocks. Therefore, the Paleocene and early Eocene paleogeographic and paleogeodynamic environments in the Eastern subregion seem quite indefinite. Apparently, at least two versions can be considered in this case: the presence of an oceanic trough (the Vetlov Formation) in the past with a primitive island arc adjoining it in the east (the Kubov Formation), or the development of oceanic or suboceanic structures, successive in time, and, later, of island-arc systems. However, any comparison with present oceanic or near-oceanic structures is still difficult to complete.

The Kilakirnun (500 m), Gailkhavilan (250 m), Alugin (more than 900 m), and the Pakhacha formations (middle Eocene to lower Miocene) belong to a rock group whose formation started in the middle Eocene after an interruption in sedimentation. A hidden disconformity may be suggested at the base of the Pakhacha Formation because the data of geological mapping, carried out in different areas, indicate that it transgressively overlies formations of different ages. In fact, all the sequences are rich in volcanic material, ash beds, and mixed rocks. Here, as in Western Kamchatka, we can observe pseudocyclicity with fine-size clastic tuffs in the upper part of elementary cyclothems. The similarity of mollusks suggests that these formations correspond to the second and third Paleogene cycles in Western Kamchatka.

THE DEVELOPMENT OF FAUNA AND FLORA IN THE PALEOGENE

The analysis of the Kamchatka Paleogene organic remains, both in its western and eastern subregions, shows that certain alternation of the marine fauna and flora assemblages went on in the sections, reflecting the evolution of ancient communities and changes in these communities depending on climatic fluctuations. As to molluscs, which were studied by V.N. Sinel'nikova and Yu.B. Gladenkov (Gladenkov *et al.*, 1991), 9 or 10 assemblages of different ages reflecting stages of their evolution can be identified in the Paleogene (Fig. 1). The most ancient of them (the Getkilnin Paleocene assemblage) is relatively poor and is characterized by small-shell fauna (Nucules and others) of the temperate type; the fauna differs strongly from that of Cretaceous assemblages. A large number of genera and species become extinct at the Gretaceous/Paleocene boundary, giving way to a new faunal type with a substantially renewed generic composition. The next, Kamchik assemblage (Late Paleocene) is marked by the appearance of some subtropical species (oysters and others) to which tropical forms (a number of gastropods, in particular) were added at the Tkapravayam level (the beginning of the lower Eocene).

The major part of the Early Eocene fauna is still poorly studied. The mollusks of the Kylan Formation (*Trochoceritium* and others), which are generally characteristic of subtropical assemblages may be assigned to this period. The middle Eocene assemblages have been more thoroughly studied. These are Snatol mollusks; their subtropical forms exhibit great diversity (135 species). Among the newly appearing species, the representatives of the *Miltha*, *Crassatella*, *Acanthocardia*, *Tivela*, *Conus*, and other genera are present here. They are succeeded by late Eocene assemblages, which are also generally subtropical (*Pitar*, *Turcicula*, *Marchallena*, and other genera), though boreal elements have already appeared as well.

In contrast, the Oligocene communities mostly comprise temperate (boreal), relatively cold-water fauna, which is assigned to a part of the Amaninian - Gakkhian regiostage, and mainly to the Utkholokian - Viventekian regiostage (in total, more than 100 forms). Representatives of the following genera are characteristic of the Oligocene: *Yoldia*, *Nuculana*, *Phacoides*, *Periploma*, and others. The Oligocene stage was a turning point in the development of malacofauna. In fact, it was the time when the core of the generic composition of Neogene assemblages was formed and when changes occurred on the species level. The early Miocene (the Kuluven-Pakhacha levels) was marked by the appearance of south boreal elements, represented in many cases by large-shelled forms (*Papyridea*, *Laternula*, and others).

Changes in molluscan assemblages recorded in the Paleogene of Kamchatka, is not only connected with the evolution of certain genera; it is also related to fauna migration. First, the migration of warm-water elements from southern latitudes (from Japan and so on) is reflected in the assemblages due to climatic warming at the beginning of the early Eocene and, later, in the middle and late Eocene, causing a northward shift of the distribution areas of tropical and subtropical assemblages.

At the same time, foraminiferal communities, both benthic and planktonic (data of V.N. Ben'yamovskii), also underwent certain changes in the Paleogene in Kamchatka. There are about five types of assemblages that characterize this period. The lowest of them, the early South Il'pinskii assemblage (the lowermost Paleocene), Rzhekhakin, is characterized by warm temperate fauna. Characteristic of the late South Il'pinskii assemblage (the Late Paleocene) is the appearance of subtropical and tropical planktonic forms. It may be called *Globorotalia* assemblage, according to the name of the dominant species. The lower Eocene is marked by the Kylan assemblage, which contains typical subtropical *Hastigerina* and *Uvigerina*. A noticeable change of assemblages was confined to the middle Eocene.

The middle Eocene Kilakirnun-Gailkhavilan assemblage comprises subtropical species with tropical elements and, according to its characteristic forms, it belongs to the *Globigerapsis-Caucasina-Plectofrondicularia* community, which is distributed in many regions of the North Pacific. Finally, the Oligocene interval is characterized by the Alugin assemblage, with the typical *Haplophragmoides-Elphidiidae-Nonionidae* community of the temperate type. At this level, a sharp change of foraminifers occurred contributing to the formation of the core of the assemblages of the Miocene boreal latitudes.

According to Chelebaeva, microscopic plant remains make it possible to identify in the Paleogene of Western Kamchatka eight characteristic flora types, corresponding to different stages of vegetation changes in the region. There are two large gaps in their successive series: at the level of the Lower-Middle Eocene stratigraphic hiatus and at the level of the Oligocene.

The earliest type is supposed to be Danian (flora of the Khulgun and Cape Zubchatyi formations) and represented by temperate flora, with predominant dentate-leaved angiosperms with medium-sized leaves, and abundant in Taxodiaceae and Platanaceae in valley communities. Late Paleocene new flora (the Sosopkhan and Kamchin Formations), being a successor of the former, differed from it in a broader spectrum of thermophilic angiosperms and a notably increased role of plants with intact-margined plate, as well as gigantophyllia (recorded in many species). All of this indicates a warmer climate, which allowed the penetration of subtropical elements. The third flora type apparently belongs to the Paleocene upper part and the Eocene lower part (the Tkapravayam Formation of the Chemurnaut Group). This is the most thermophilic flora in Kamchatka, comprising many forms of the early Ravenian paratropical flora of Alaska. Unlike the latter, it is of a more humid type and taxonomically it is a more clearly pronounced successor of the Paleocene floras. Palms, Myrtaceae, Leguminosae plants are typical components of the Tkapravayam flora.

Higher, after some time interval, follows the resembling subtropical, mesophilic Ust'-Anadyrka flora, which preserves forms typical of the optimum Early-Middle Eocene North American floras (*Calkinsia* and *Cycas*). It was replaced by the early Snatolian flora, corresponding apparently to a short-term climatic deterioration in the middle of the middle Eocene, which was marked by a sharp increase in the role of Taxodiaceae and a decrease in the role of intact-margined plants. The upper part of the middle Eocene (Bartonian Stage) is characterized by uniform flora of the upper Snatolian and Kovachinan regiostages, abundant in diverse thermophilic conifers (*Pinus*, *Cunninghamia*, and *Sciadopitys*), which reflect a nearly subtropical climate and an intensely dissected young topography, probably volcanic. The Kovachinan level is characterized by the wide distribution of intact-margined *Macclintokia* leaves in marine coastal facies.

The upper Eocene stage is represented by the mesophilic warm temperate Irgirnin flora, which, on the one hand, seems to be a successor the Paleocene-Eocene temperate floras and, on the other, shows a renewal at the specific and generic levels, in particular an increase in the diversity and significance of beeches and birches. The Krutogorovo flora may be referred to the late Paleogene floras; however, its age is provisionally defined as transitional Oligocene to early Miocene. The composition of this warm temperate mesophilic flora has little in common with the Paleocene-Eocene floras of the same type. It has preserved few genera and species characteristic of the latter (*Woodwardia*, *Osmunda*, *Glyptostrobus*).

Incidentally, neritic diatomic flora became rather widely distributed in this area beginning with the Oligocene (Gladenkov, 1991). On the whole, the analysis of Paleogene faunal and floral communities in this region has made clear a definite trend in the change of their ecological types. Thus, the warmest (up to tropical) and most diverse faunal and floral assemblages were confined to the level of the end of the Paleocene-beginning of the Eocene. Thermophilic (subtropical) sea and land associations have been recorded in the Middle-Upper Eocene. Almost simultaneously and rather sharply, they were replaced by warm temperate and relatively cold-loving elements of the Oligocene stage.

Simultaneously, a certain change in the systematic composition, coinciding in time, took place in these communities. It affected both the specific and generic composition that resulted from the summary effect of the evolution of the assemblages and, to a certain extent, from the migration of various elements in certain moments of the past. In early Eocene time, such rearrangements took place both in the sea and on the land. They were also manifested in the Middle-Late Eocene stage. Finally, sharp changes in the composition of complexes occurred in the Oligocene (probably, they began in the late Eocene), making possible the formation of the starting associations that developed in Neogene time.

CLIMATIC EVENTS

The available data on fauna and flora indicate a general relative cooling of the climate during the Paleogene. Certain climatic fluctuations were also superimposed upon this tendency. Three major warmings of the climate can be recognized in the Paleogene history of Kamchatka. The first (early) and most intense warming occurred at the end of the Paleocene and the beginning of the early Eocene; it was clearly reflected in tropical molluscan and floral assemblages from the Kamchatka sections. The second and third climatic optima, somewhat weaker than the first, were manifested in the middle and late Eocene. They were marked by marine faunal assemblages and by flora of the subtropical type. The relative cooling began at the beginning of the Oligocene (possibly, at the end of the

Eocene). It was reflected in faunal boreal communities and in rather cold floral assemblages. It is to be noted that short-term rises in temperature are believed to have occurred during the Oligocene cooling. The early Miocene was characterized by a warmer (compared to the Oligocene) temperate climate. At this time, south boreal species of benthic groups, including large-shelled mollusks, appeared at the latitude of Kamchatka.

Judging from the materials on flora, the subtropical and paratropical climate of the Paleocene was also humid, while the rise in temperature (which occurred in the middle Eocene) was accompanied by aridization. The warm humid climate was re-established by the end of the Eocene and changed to the temperate-cool in the Oligocene (Chelebaeva and Brattseva, 1985).

The presence of floating pebbles, which in some cases may be regarded as ice rafting material, at certain levels of the Paleogene sequence and gennoish concretions (Kraevaya *et al.*, 1987) may serve as indicators of climatic changes (cooling, in particular). The appearance of such concretions in the Snatolian middle Eocene rocks might relate to the change of the early Eocene warm humid climate to the middle Eocene climate, with periodical cold seasons. The most notable gennoish concentrations are recorded in the Oligocene at the Gakkhian and, particularly, at the Utkholokian-Viventekian levels; that may indicate a considerable cooling in Oligocene time. Major floating pebble concentrations are assigned to the Oligocene (the Alugin level); incidentally, at the bottom of the Utcholok Formation, the pebbles are associated with boulders, the appearance of which might reflect a period of relative cooling.

It is interesting that the middle Eocene warming (Chelebaeva and Brattseva, 1985) coincided with the beginning of the sea transgression in Kamchatka, whereas the Oligocene fall of temperature may be correlated with the transgressive-regressive phase. It shows that we cannot unambiguously define the interdependency of various geological phenomena (in particular, warmings did not always coincide with the beginning of transgressions, as often occurred during the Neogene in Kamchatka). Probably, subglobal geological events (for instance, ocean level eustatic variations and climatic fluctuations) are associated with local ones, which are due to local causes (for instance, transgressions in Kamchatka could have resulted from tectonic movements).

CONCLUSION

Many difficulties are encountered in the study of Paleogene sections in Kamchatka. First, they are due to the intricate tectonic structure of the region. This is particularly true of the eastern regions (the Vostochnyi Range, the Kronotskii Peninsula, the eastern part of Karaginskii Island, and the Goven Peninsula), which are marked by widespread Eocene (?) and younger imbricate overthrusts, while ancient sequences are crumpled into the steep folds of the eastern vergence. On the other hand, the difficulties may arise from the facies diversity of Paleogene beds in different structural zones.

The analysis of the available material allows us to speak of the generally cyclic development of the region in Paleogene time, which was reflected in volcano-sedimentary cycles and in the stages of organic development. It also shows that sedimentation took place in relatively shallow seas (maybe with the exception of eastern parts of the region). At the same time, we may speak of the regime of a peculiar unstable orogenic belt in the western and central parts of Kamchatka and the island-arc oceanic regime (?) in the southern part of Kamchatka in Paleocene to early Eocene time, as well as about the existence of eastern and western volcanic zones in the Paleocene and Eocene, their subsequent disappearance, and the concentration of volcanism in the central and southern parts of Kamchatka.

The data accumulated indicate both similarity and significant differences in the paleogeographic environments of the Kamchatka Paleogene as compared with recently active convergent continental margins. The material presented in this paper, confirms the complex evolution of the Paleogene continental margin (with reference to Kamchatka, which acquired its features only in the late Cenozoic) (Shantser, 1979).

The data suggest a certain tendency in the change of ancient marine associations (mollusks and foraminifers). The change of their genera and species occurred simultaneously (at least within the regiostages and their parts) though Paleocene and lower Eocene assemblages need to be studied more carefully.

It should be emphasized that the Oligocene was (to a considerable extent) a turning point in the development of marine biota because the assemblages formed in the Oligocene differed considerably from ancient ones and, in fact, composed the core of the associations developed in the Miocene.

The formation of biocenoses in the Kamchatka shelf zones was strongly affected by ocean level eustatic variations and climatic fluctuations. In particular, during the cooling period at the latitude of Kamchatka, many migrants moved there from southern latitudes.

The turning points in the development of the terrestrial floras correspond to past climatic fluctuations. However, in marking successive changes of floral communities in the Paleogene, we should emphasize that their most substantial rearrangement falls at the end of the Eocene and the beginning of the Oligocene; this brought about the formation of a new type of plant association, the development of which continued in Neogene time. Thus, broad faunal and floral changes, which occurred in different ecosystems of the Kamchatka Paleogene, appear to be generally synchronous.

Investigations of the Paleogene marine fauna in Kamchatka show that the Kamchatka assemblages had clear connections and interchanges with both North American and Japanese communities. It seems that

Paleocene (?) and early-middle Eocene shelf assemblages within the North Pacific were quite similar (being on the whole thermophilic) and, in fact, embraced all the shelf zones of the region. The connections between east Asian and west American associations became weaker in late Eocene time. This problem must be studied additionally because changes of the paleogeographic and paleoclimatic conditions might provoke fauna asymmetric migrations and disturb the bilateral biota interchange in the Asian and North American shelves. The lack of comparative material does not allow us to speak about prohoreses of Oligocene time, though we may suggest that there was a rather intensive fauna interchange within the North Pacific shallow-water zones (probably, the migration of Asian forms toward the North American shelf prevailed).

We should particularly emphasize that the similarity of Oligocene diatomic assemblages of Kamchatka and other regions of the North Pacific with coeval assemblages of the North Atlantic region may serve as a confirmation of the connection between the North Atlantic, Arctic, and Pacific oceans. On the whole, we should draw attention to the observed coincidence of biotic and abiotic events in the period under consideration. This phenomenon must be studied specially, as the recorded synchroneity of certain natural events does not always directly indicate their cause and effect relations.

REFERENCES

Budantsev, L.Yu., Phytostratigraphy of the Continental Paleogene in West Kamchatka, *Materialy po Stratigrafii i Paleogeografii Vostochnoi Azii*, DVO AN SSSR, Vladivostok, 1984, pp. 53 - 58.

Chelebaeva, A.I. and Brattseva, G.M., On the Paleogene Climatostratigraphy According to Paleofloras (North Pacific Region and Adjacent Areas), *Korrelyatsiya Kainozoiskikh Tolshch Dal'nego Vostoka*, GIN AN SSSR, Moscow, 1985, Part 2, p. 157 - 209.

Chelebaeva, A.I. and Shantser, A.E., New Data on the Early Paleogene of West Kamchatka, *Litologiya i Stratigrafiya Mezozoya i Kainozoya Vostochnykh Raionov SSSR*, Nauka, Moscow, pp. 135 - 148.

Chekhovich, V.D., Bogdanov, N.A., Kravchenko-Berezhnoi, I.R., Averina, G.Yu., Gladenkov, A.Yu., and Til'man, S.M., *Geologiya Zapadnoi Chasti Beringovomor'ya* (Geology of the Western Part of the Bering Sea Region), Nauka, Moscow, 1990.

Gladenkov, Yu.B. and Shantser, A.E., Stratigraphy and the Paleogeographic Background of the Correlation of Neogene Geological Events in the Kamchatka Region, *Osadochnaya Obolochka Zemli v Prostranstve i vo Vremeni*, Papers of Soviet Geologists at the XXVIII Session of the International Geological Congress, Nauka, Moscow, 1989, pp. 127 - 134.

Gladenkov, Yu.B., Sinel'nikova, V.N., Shantser, A.E., *et al.*, *Eotsen Zapadnoi Kamchatki* (The Eocene of West Kamchatka), Nauka, Moscow, 1991.

Gladenkov, Yu.B., Bagdasaryan, G.P., Ben'yamovskii, V.N., *et al.*, Plankton in the Paleogene of the Il'pinskii Peninsula (Koryak Highland), *Izv. AN SSSR, Ser. Geol.*, 1988, no. 10, pp. 85 - 90.

Gladenkov, Yu.B., Sinel'nikova, V.N., Chelebaeva, A.I., and Shantser, A.E., Problems of the Paleogene of the Tigil' Region in West Kamchatka, *Korrelyatsiya Kainozoiskikh Tolshch Dal'nego Vostoka*, GIN AN SSSR, Moscow, 1985, P. 1, pp. 45 - 55.

Gladenkov, Yu.B., Chelebaeva, A.I., and Shantser, A.E., New Data on the Paleogene of the Gulf of Shelekhov Marine Coast (Northwestern Kamchatka), *Izv. AN SSSR, Ser. Geol.*, 1990, no. 6, pp. 7 - 87.

Gladenkov, A.Yu., *Stratigrafiya Kainozoya Kamchatskogo Regiona po Diatomeyam (Oligotsenovyi i Pogranichnyi Pliotsen-Chetvertichnyi Urovni)*, Cenozoic Stratigraphy of the Kamchatka Region on the Basis of Diatoms (Oligocene and the Pliocene-Quaternary Boundary Levels), Cand. Sci. (Geol.-Min.) Dissertation, Institute of Lithosphere, Moscow, 1991.

Khubunaya, S.A., *Formatsionnaya Prinadlezhnost' Melovykh i Paleogenovykh Bazal'tov Kronotskogo Poluostrova i Eostenovykh Bazal'tov Kamchatskogo Poluostrova (Vostochnaya Kamchatka) po Geokhimicheskim Priznakam* (The Formation of Cretaceous and Paleogene Basalts in the Kronotskii Peninsula and Eocene Basalts in the Kamchatka Peninsula (East Kamchatka) According to Geochemical Features), Cand. Sci. (Geol.-Min.) Dissertation, Moscow, 1984.

Kraevaya, T.S., Chelebaeva, A.I., and Shantser, A.E., Gennoish Morphological Types in Cenozoic Strata of Kamchatka and Their Possible Climatostratigraphic Significance, *Litologiya i Poleznye Iskopaemye*, 1987, no. 6, pp. 131 - 135.

Serova, M.Ya., Fot'yanova, L.I., and Fradkina, A.F., Marine and Continental Deposits in Northwestern Kamchatka, *Kainozoi Dal'nego Vostoka*, DVO AN SSSR, Vladivostok, 1989, pp. 186 -189.

Shainyan, S.Kh., Stratigrapic Subdivisions of the Cenozoic on the North Okhotsk Shelf, Abstracts of Papers, the 3rd Pacific School on Marine Geology, *Geophysics and Geochemistry*, Part 2, p. 76.

Shantser, A.E., *Kainozoiskoe Razvitie Kamchatki: Formirovanie i Destruktsiya Nestabil'nykh Orogenicheskikh Podnyatii*, (Cenozoic Development of Kamchatka: Formation and Destruction of Unstable Orogenic Elevations), Nauka, Moscow, 1987.

Shantser, A.E., Kamchatka Block Structure and the Location of Tectono-Magmatic Structures Depending on the Late Cenozoic Movements of Blocks, *Byull. Vulkanologicheskoi Stantsii*, 1979, no. 57, pp. 25 - 36.

Shantser, A.E., Cyclicity of Geological Development of Kamchatka in the Late Cenozoic, *Genezis Osadkov i Fundamental'nye Problemy Litologii*, Nauka, Moscow, 1979, pp. 217 - 221.

Stratigraphy and Geological Correlation, Vol. 1, No. 1, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 20, No. 2, 1993, pp. 109 - 117.
Original Russian Edition Copyright © 1993 by Stratigrafiya. Geologicheskaya Korrelyatsiya, Repin, Polubotko.
English Translation Copyright © 1993 by Interperiodica.

Zonal Subdivision of the Upper Toarcian in the Northeast of Russia

Yu. S. Repin and I. V. Polubotko

Russian Research Institute of Petroleum Exploration, Liteinyi Pr. 39, St. Petersburg, 191104, Russia

Manuscript received May 5, 1991

Abstract – A continuous section of Lower and Middle Jurassic strata containing abundant ammonite remains, exposed in the Letnyaya River (in the basin of the Berezovka River, a right tributary of the Kolyma River) and enclosed between the upper Toarcian *Pseudolioceras rosenkrantzi* Zone and the *Pseudolioceras beyrichi* Zone (the base of the Aalenian), has yielded two more ammonite zones: *Pseudolioceras danilovi* and *Pseudolioceras paracompactile*. They are believed to add to the upper Toarcian zonal scale in northeastern Asia. A description of three new species of Pseudolioceras is given, including the index species of the proposed zones, and *Pseudolioceras evolutum* Repin, *sp. nov.*

Studies of the Jurassic biostratigraphy of northeastern Asia have proved particularly successful in the last quarter of the century. Classified formerly among the least studied, the region now has zonal scales, which (in terms of substantiation and details available) approach those of the stratotype regions of Jurassic stages, with century-long histories of investigation behind them (*Stratigrafiya...*, 1976; *Resheniya...*, 1978; *Resheniya...*, 1981; *Zony...*, 1982; Polubotko and Repin, 1988).

The scale of ammonite zones of the Toarcian Stage in northeastern Russia shows more units than the other Lower and Middle Jurassic stages. However, six out of the seven local zones in the stage refer to the lower Toarcian. The upper Toarcian is represented in the latest scales by a single *Pseudolioceras rozenkrantzi* Zone correlated with the entire range of the upper Toarcian, which in the general scale is subdivided into three zones (Fig. 1). A degree of convention involved in such correlation, as well as the unclear range and paleontologic record of the upper Toarcian in northeastern Asia (or its presence in some regions), has long been attracting the attention of the researchers, posing a number of questions that have not yet been resolved.

There are several reasons for the problem of the upper Toarcian in northeastern Russia and northern Siberia. First, Toarcian-Aalenian ammonite levels (zones) have not been recorded in a single continuous section, and there are frequent stratigraphic gaps at that level. Second, climate is known to have been changing in the late Toarcian: following the Early Toarcian climatic optimum, boreal basins' waters must have cooled on a global scale, leading to a higher differentiation between the Tethyan and boreal faunas. Boreal molluscan communities abruptly lost much of their diversity, while endemic ammonite phyletic lines emerged. Ammonites had retained a single genus of Pseudolioceras, making it more difficult to directly correlate them with the sequence of ammonite levels in the Toarcian standard, whereas Pseudolioceras genus is but subsidiary in importance.

The problem is further confused by poor correlation between the east of the Siberian platform and the northeast of Russia. The reason is higher biogeographic differentiation that produced in the late Toarcian two provincial biochores: Yakut and Kolyma.

Subdivision of the Toarcian into zones has been elaborated on the sequences in the central Omolon massif, in the upper reaches of the Levyi Kedon River (Dagis and Dagis 1965; Polubotko and Repin 1966). Leaving out any in-depth discussion here of the validity or nomenclature variations in the stratigraphic units of this scale, whose latest version is shown in Fig. 1, note that in this section, coming above the *Pseudolioceras (P.) rosenkrantzi* Zone with erosion and stratigraphic hiatus are strata of siltstones and sandstones bearing in its lower portion fossils of *Pseudolioceras (Tugurites) fastigatum West.*, an index species of the lower Bajocian zone in the regional scale of the Jurassic of the Northeast of the former Soviet Union (*Resheniya...*, 1978). The upper Toarcian there is represented by a single P*seudolioceras rosenkrantzi* Zone, and the remains of the species occur as far as the gap's boundary.

Aalenian zones in eastern, and northeastern Russia and northern Siberia have been recorded in the sequences of the Torom basin in Western Priokhot'e (Sei and Kalacheva, 1980). In the coast of the Tugur Bay, Aalenian strata, beginning with the beds with *Pseudolioceras (P.) beyrichi*, which overlie the strata of the lower Toarcian *Zugodactylites monestieri* Zone. The entire upper Toarcian, together with the *Peronoceras spinatum* Zone of the lower Toarcian, is missing from the section (Fig. 1).

Although critical for the zonal stratigraphy, the Jurassic section in the Viliga River, northern Priokhot'e

		Standard zones	Zones of the Russian Northeast	Levyi Kedon River	Tugur Bay	Viliga River	Markha River
AALENIAN	Upper	*Graphoceras concavum*	*Pseudolioceras (Tugurites) tugurense*		*Pseudolioceras tugurense*	*Pseudolioceras tugurense*	
		Ludwigia murchisonae					
	Lower	*Leioceras opalinum*	*Pseudolioceras (Pseudolioceras) maclintocki*		*Pseudolioceras maclintocki*	*Pseudolioceras maclintocki*	
			Pseudolioceras (Pseudolioceras) beyrichi		*Pseudolioceras beyrichi*	*Pseudolioceras beyrichi*	?.....
TOARCIAN	Upper	*Dumortieria levesquei*	*Pseudolioceras (Pseudolioceras) paracompactile*			Basalts	*Pseudolioceras (Pseudolioceras) alienum*
		Grammoceras thouarsense	*Pseudolioceras (Pseudolioceras) danilovi*				
		Haugia variabilis	*Pseudolioceras (Pseudolioceras) rosenkrantzi*	*Pseudo-lioceras rosenkrantzi*			
	Lower	*Hildoceras bifrons*	*Peronoceras spinatum*	*Peronoceras spinatum*			
			Zugodactylites monestieri	*Zugodactylites monestieri*	*Zugodactylites monestieri*	*Zugodactylites monestieri*	*Zugodactylites monestieri*
			Dactylioceras athleticum	*Dactylioceras athleticum*		*Dactylioceras athleticum*	*Dactylioceras athleticum*
		Harpoceras falciferum	*Harpoceras falciferum*	*Harpoceras falciferum*		*Harpoceras falciferum*	*Harpoceras falciferum*
			Eleganticeras elegantulum	*Eleganticeras elegantulum*			*Eleganticeras elegantulum*
		Dactylioceras tenuicostatum	*Tiltoniceras propinquum*	*Tiltoniceras propinquum*			

Fig. 1. *Subdivision of Lower/Middle and Jurassic strata in the northeast of Asia into zones.*

(Polubotko and Repin, 1974; *Stratigrafiya...*, 1976), fails to shed any light on the upper Toarcian problem, since basalt lava breccia (200 m) that occurs between the lower Toarcian *Zugodactylites monestieri* Zone and the lower Aalenian *Pseudolioceras beyrichi* Zone, contains in some tuff layers only rostra of unidentifiable belemnites.

Thus, of particular interest is a continuous section of Lower/Middle Jurassic boundary strata cropping out in the middle course of the Berezovka, a right tributary of the Kolyma River (Fig. 2). We have studied the section in 1987.

The section is a sequence of clayey and silty rocks, abundant in small clayey carbonate nodules. Transi-

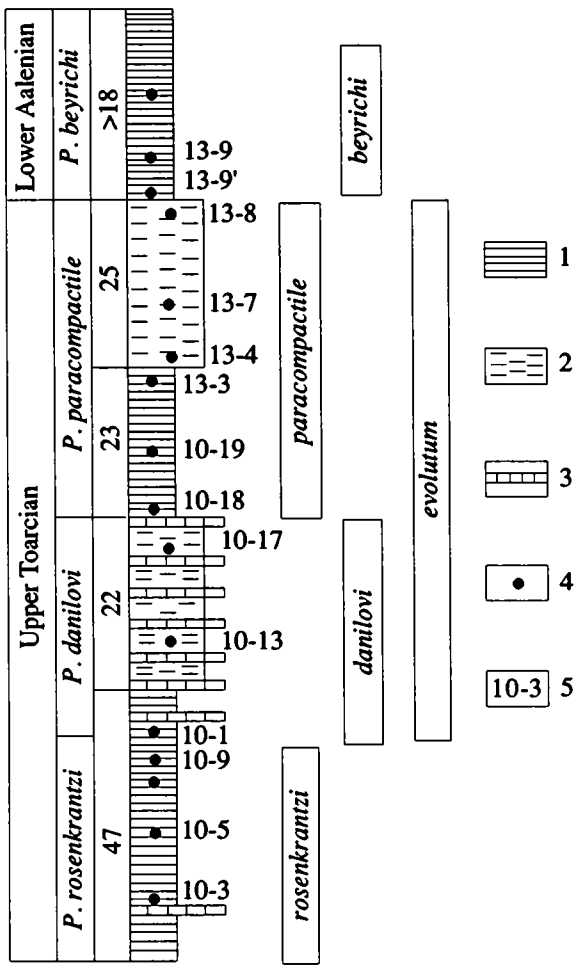


Fig. 2. *A stratigraphic column and a diagram showing distribution of Pseudolioceras species over the section of Lower and Middle Jurassic strata in the Berezovka River basin.*

1 – mudstones; 2 – siltstones; 3 – interbeds of clayey limestone; 4 – carbonate nodules; 5 – sampling sites and field numbers of samples with fauna.

tions between the identified members are vague and the boundaries are to some extent arbitrary. From the base upward the section is composed of:

(1) Dark-grey mudstones, splinty and with concentric parting, reddish-brown on the weathered surface. They contain rare lens-shaped 5 - 15-cm interbeds of clayey limestones. At a distance of 5 - 8 m from the base of the bed, there are many clayey-carbonate nodules, varying in shape from flattened, pan-like to rounded, 3 - 5 cm to 10 - 12 cm across. At the top of the member, there is a persistent, up to 40 cm thick layer of clayey limestone. The apparent thickness is 47 m.

(2) Siltstones and clayey siltstones, dark-grey, splinty. They contain abundant interbeds of clayey limestone, 10, 15, and 20 cm thick, occurring every 1.0, 1.5, 2.0 and 3.0 m. Present all over the member are clayey carbonate nodules, usually ball-shaped and rarely more than 3 - 5 cm across. The thickness is 22 m.

(3) Clayey siltstones and mudstones, dark-grey, with abundant clayey carbonate nodules, from 3 - 4 to 10 cm across, particularly abundant in the middle of the member. The thickness is 22 - 24 m.

(4) Siltstones and clayey siltstones similar to those of bed 2, but virtually devoid of interbeds of clayey limestone, showing fewer clayey carbonate nodules. The thickness is about 25 m.

(5) Mudstones and clayey siltstones, with rare lenses of clayey limestone, 1.5 - 3.0 m long and up to 30 - 40 cm thick, with a brown surface. Nodules do not form any notable clusters. The apparent thickness is about 18 m.

The total thickness of the section is 135 m.

The section shows uniformly distributed ammonite fossils confined generally to the nodules, and rarely to interbeds of clayey limestone. Ammonites belong to a single genus of *Pseudolioceras*. About 4000 specimens have been collected from the section, of which only one was a specimen of *Phylloceras heterophyllum* (Sow.) (sample 10-3). Small bivalves were found in the nodules from only one bed (sample 10-13) and identified as *Arctotis aff marchaensis (Petr.), Mytiloceramus (Lenoceramus) sp.*, and *Camptonectes sp.*

The uniform fine-grained rocks, their large thickness, and the assemblage of fossils are suggestive of the fairly deep-water sedimentary environment and of no apparent and substantial stratigraphic gaps.

The consecutive distribution of *Pseudolioceras* species over the section identifies the following zones (lones). The lower 37 meters of member 1 are characterized by *Pseudolioceras (P.) rosenkrantzi* A. Dagis, and are referred to the eponymous zone of the upper Toarcian. The upper 10 meters of member 1 and the entire member 2 contain another two species: *Pseudolioceras (P.) danilovi* Repin, *sp. nov.*, and *Pseudolioceras (P.) evolutum* Repin *sp. nov.* and are identified as the *Pseudolioceras (P.) danilovi* Zone. Member 3 and the lower portion of member 4 are characterized by the remains of *Pseudolioceras (P.) evolutum* Repin, *sp. nov.*, and *Pseudolioceras (P.) paracompactile* Repin, *sp. nov.* This part of the section is referred to the *Pseudolioceras (P.) paracompactile* Zone, together with the upper part of member 4, arbitrarily included in here as it yielded no identifiable ammonites. A change in Pseudolioceras species takes place at the base of member 5.

Member 5 yielded remains of *Pseudolioceras (P.) beyrichi (Schloenb.)*, an index species of the lower Aalenian zone, suggesting this bed may belong to the Middle Jurassic.

Hence, in the above section between the *Pseudolioceras rosenkrantzi* and *Pseudolioceras beyrichi* Zones, there are two discrete levels that are proposed as two new biostratigraphic zones of the upper Toarcian. Apart from the position of these two zones in the section beneath the *Pseudolioceras beyrichi* Zone, which is the lowermost Aalenian ammonite level in northeastern Asia, another evidence of the upper Toarcian origin is

the close affinity of the local species of *Pseudolioceras (P.) paracompactile sp. nov.* and *Pseudolioceras (P.) compactile* (Simpson), which is a common element of the late Toarcian ammonite communities in Western Europe. It is particularly frequent in the middle part of the upper Toarcian (*Grammoceras thouarsense* Zone), although there is evidence of its presence in the uppermost Toarcian (Dietl and Etzold 1977; Plate I, Fig. 3).

We regard the east Asian *P. paracompactile sp. nov.* as a vicariant of the European *P. compactile* (Simpson), assuming that they must have a close stratigraphic range.

The index species of the upper Toarcian lower zone in the Northeast of the former Soviet Union, *Pseudolioceras (P.) rosenkrantzi* A. Dagis appears to be fairly close to the European *Pseudolioceras (P.) württembergica* (Denckmann), so that they can also be considered as vicarious species. The stratigraphic position of the latter species is not quite clear, although it hardly falls out of the range of the lower half of the upper Toarcian. So far, it is impossible to directly correlate the upper Toarcian zones in the northeast Russia, which are identified by a change in the species of the endemic Pseudolioceras line, and the standard zones and subzones, established on the basis of the succession of Late Toarcian genera of *Haugia, Grammoceras, Pseudogrammoceras, Phlyseogrammoceras, Dumortieria* and *Pleydellia.*

What follows below is the description of three new species of Pseudolioceras collected from the section in the Berezovka basin. The collection is stored in the museum of the Russian Research Institute of Petroleum Exploration (VNIGRI) in St. Petersburg, item No. 837.

Order Ammonitida

Suborder Ammonitina

Family Hildoceratidae Hyatt, 1867

Subfamily Harpoceratinae Neumayr, 1875

Genus *Pseudolioceras* Buckman, 1889

Subgenus *Pseudolioceras* Buckman, 1889

Type species: *Ammonites compactilis* Simpson, 1885. The Toarcian Stage of England.

Pseudolioceras (Pseudolioceras) danilovi Repin *sp. nov.*

Plate I, Figs. 5 and 6; Plate II, Figs. 4, 5, 6, 8, 9; Plate III, Figs. 4, 7, 8.

Named after geologist V.G. Danilov.

Holotype. Specimen 52/837, Plate II, Fig. 4. Upper Toarcian, the *Pseudolioceras danilovi* Zone in the Northeast of the former Soviet Union.

Material. About 50 complete cores and a few dozens of casts and fragments of shells.

Description. A medium-size shell (a maximum diameter of 70 mm) shows involute and flattened whorls. The side surfaces of the shells 30 - 35 mm in diameter are only slightly convex, with the thickest portion of the whorl coming above its middle. With the growth of the shell, the most convex portion is shifted to the middle of the whorl. Side surfaces smoothly but rapidly grade into a rounded and narrow, keel-shaped sharp venter. The conjunction of the sides with the umbilicus wall is rounded in shells with a diameter of 15 - 20 mm, becoming more angular farther. The umbilicus is narrow with sheer low walls. The whorls sections are high, elliptical, with the height twice the thickness of the whorl.

Typical of this species is a long development stage of thin, striated, sickle-shaped singular ribs, which continues until the shell is 30 - 35 mm across; however, there are a few specimens (Plate II, Fig. 9) whose ribs, although thin, are quite apparent even in 20-mm-diameter shells. Adult specimens show numerous relatively thin sickle-shaped ribs, indistinct in the lower half of the whorl and pronounced in the upper one. The width of triangular rounded low ribs and that of the spaces between them are approximately equal. Shells from 40 up to 60 mm in diameter have from 38 to 46 ribs, i.e., rib index[1] varies between 0.7 and 0.9.

Comparison. Among other species of the *Pseudolioceras* genus, the described one distinctly stands apart due to its long stage of weakly striated ribs, flattened lateral sides and numerous fine ribs. At the stage of weakly striated ribs, *P. danilovi* approaches, in terms of the sculpture, *Pseudolioceras lectum* (Simpson) (Buckman 1930; Plate 43, Figs. 1 - 3), but is different in its flattened whorls, with their maximum thickness positioned above the middle of the whorl, and a narrower umbilicus.

Range. The Kolyma River basin (Northeast of Russia), the Berezovka River; the upper Toarcian, the *Pseudolioceras danilovi* Zone.

Pseudolioceras (Pseudolioceras) paracompactile Repin, *sp. nov.*

Plate II, Figs. 2, 3, 7; Plate III, Fig. 9.

Pseudolioceras compactile (Simps.): (Repin, 1968; Plate 53, Fig. 1, but not Figs. 2 or 3).

The name by comparison with the P. *compactile* species.

Holotype. Specimen 131/837, Plate II, Fig. 3. The upper Toarcian, the *Pseudolioceras paracompactile* Zone, the Northeast of Russia.

Material. About 100 well-preserved cores and up to 300 casts and shell fragments.

Description. Medium-size shells (maximum diameter up to 60 mm) have whorls, rapidly gaining in height and involute. High-oval whorls cover up 80 - 85% of the preceding ones. The venter is rounded and narrow; it grades smoothly and rapidly into slightly venter-tilted lateral sides. The latter are weakly but distinctly convex. The umbilicus is moderately narrow (umbilicus to diameter ratio is 0.16 : 0.18). The umbilical wall is vertical, meeting the side surface at a right angle. The whorl is thickest at the middle.

[1] The ratio of the number of a whorl's ribs to the shell's diameter.

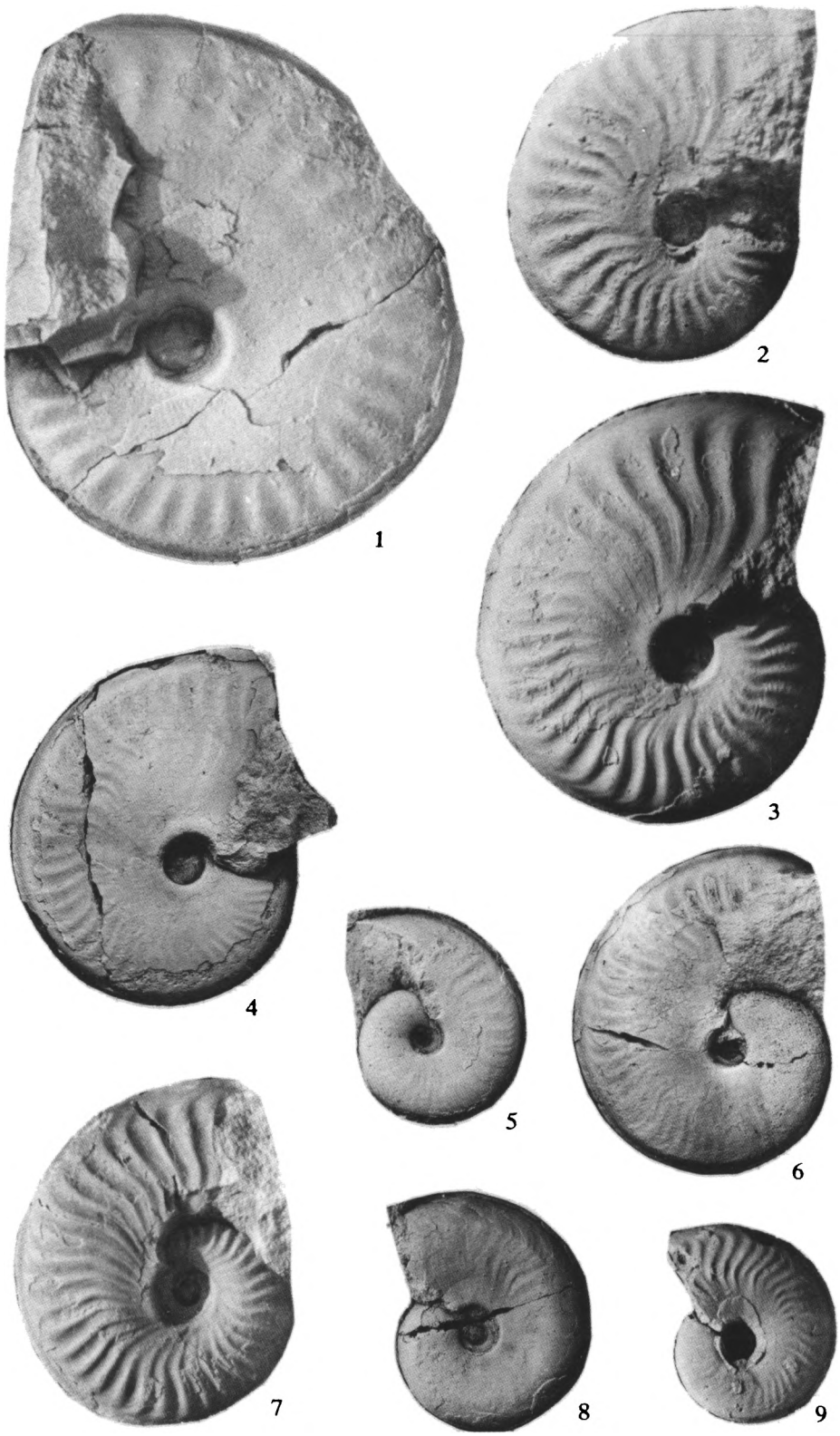


Plate I. *1 - 4 – Pseudolioceras (Pseudolioceras) rosenkrantzi A. Dagis, 1 – sample 10-7 (37/837), a phragmocone, a side view, involutoconch, 2 – sample 10-6 (31/837), a side view, evolutoconch, 3 – sample 10-7 (42/837), a side view, 4 – sample 10-3 (43/837), phragmocone, a side view. P. rosenkrantzi Zone. 5, 6 – Pseudolioceras (Pseudolioceras) danilovi Repin, sp. nov. 5 – sample 10-1 (44/837), a side view, 6 – sample 10-1 (45/837), a side view. P. danilovi Zone.*

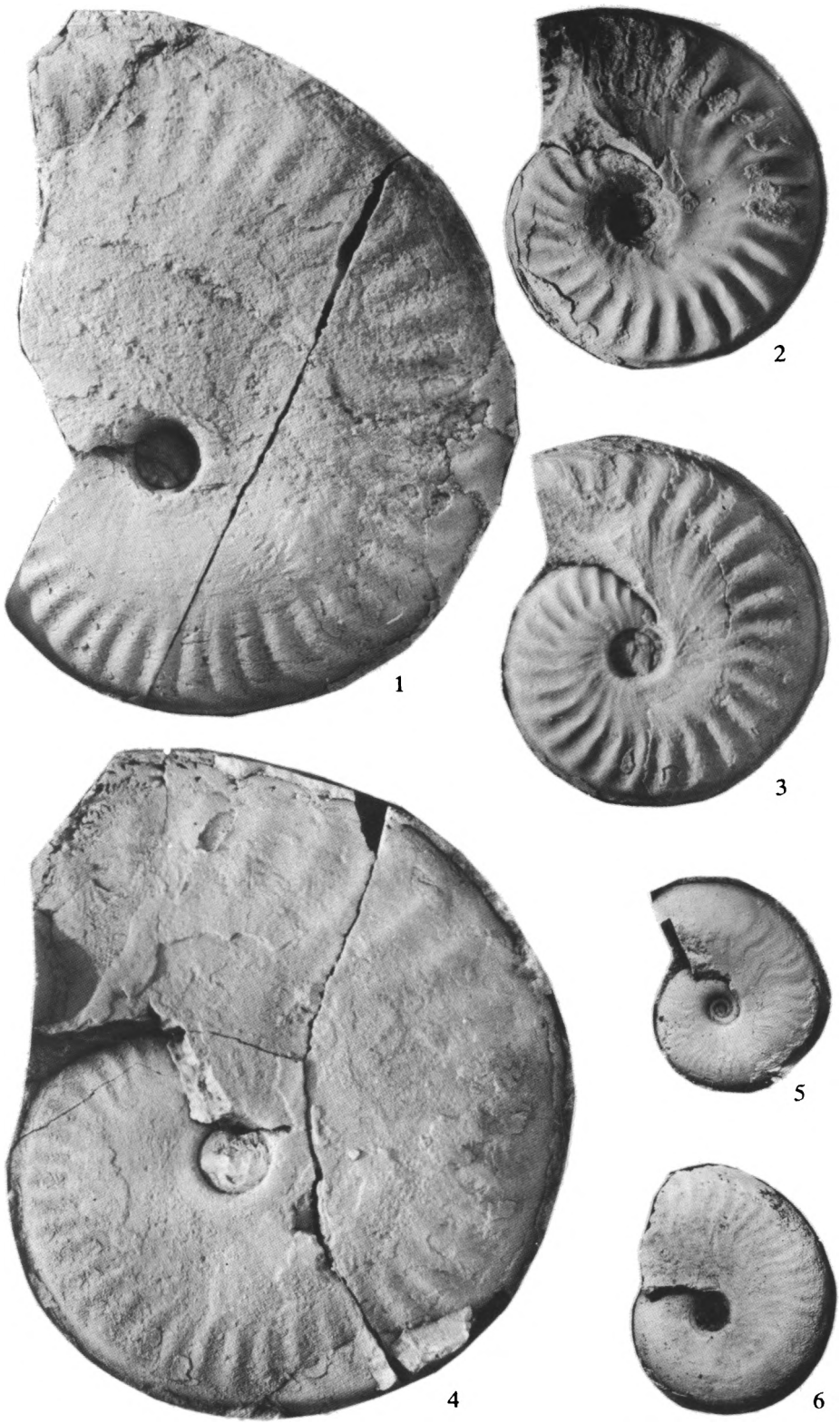


Plate II. *1 – Pseudolioceras (Pseudolioceras) rosenkrantzi A. Dagis. Sample 10-4 (61/837), a phragmocone with a beginning of the body chamber. P. rosenkrantzi Zone. 2, 3, 7 – Pseudolioceras (Pseudolioceras) paracompactile Repin, sp. nov. 2 – sample 10-18 (46/837), a side view; 3 – sample 10-19 (131/837), holotype, a side view; 7 – sample 10-19 (47/837), a side view. P. paracompactile Zone. 4, 5, 6, 8, 9 – Pseudolioceras (Pseudolioceras) danilovi Repin, sp. nov. 4 – sample 10-14 (52/837), holotype, a side view; 5 –sample 10-1 (48/837), a side view; 6 – sample 10-1 (49/837), a side view; 8 – sample 10-1 (50/837), a side view; 9 – sample 10-11 (51/837), a side view. P danilovi Zone.*

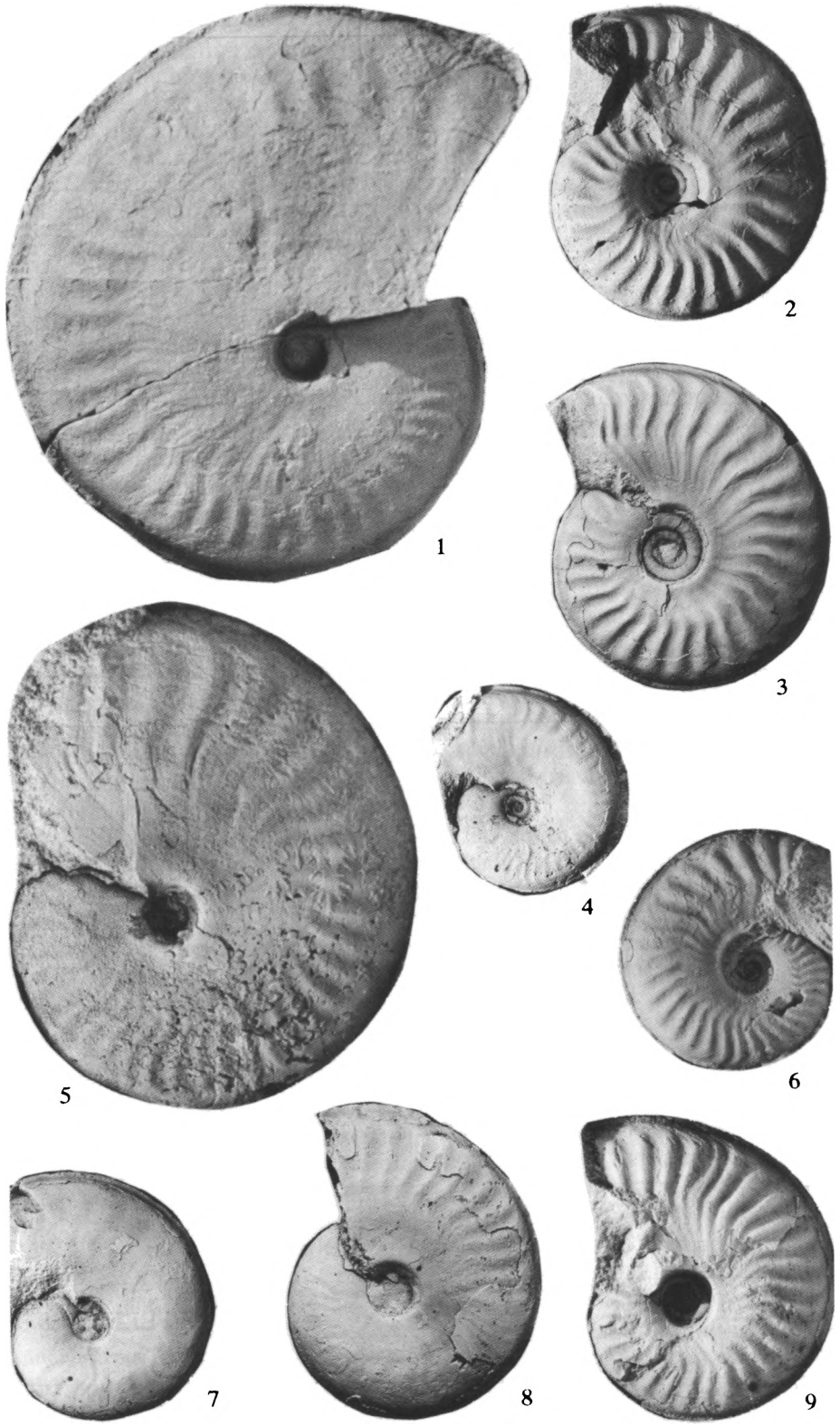


Plate III. *1, 5 – Pseudolioceras (Pseudolioceras) beyrichi (Schloenbach). 1 – sample 13-9' (53/837), a side view; sample 13-9 (54/837), a side view. P. beyrichi Zone. 2, 3, 6 – Pseudolioceras (Pseudolioceras) evolutum Repin, sp. nov. 2 – sample 10-13 (55/837), a side view. P. danilovi Zone. 3 – sample 10-19 (143/837), holotype, a side view. P. paracompactile Zone. 6 – sample 10-18 (56/837), a side view. P. paracompactile Zone. 4, 7, 8 – Pseudolioceras (Pseudolioceras) danilovi Repin, sp. nov. 4 – sample 10-11 (57/837), a side view; 7 – sample 10-11 (58/837); 8 – sample 10-12 (59/837), a side view. P. danilovi Zone. 9 – Pseudolioceras (Pseudolioceras) paracompactile Repin, sp. nov. Sample 10-18 (60/837), a side view. P. paracompactile Zone.*

The sides are covered by simple sickle-shaped ribs, starting at the inner edge of the whorl, where they are but vaguely distinct. In the upper half, ribs are higher and broader, with similarly expanding inter-rib spaces which are slightly wider than the ribs. Closer to the venter, ribs are smoother. The keel is sharp and low.

Comparison and remarks. As it follows from the name of the new species, it is closest to the Pseudolioceras genus' type *P. compactile* (Simpson). Note that although *P. compactile* has been chosen as the type of the *Pseudolioceras* genus, its specific characteristic is yet incomplete, and various authors differ in their treatment of the range of P. compactile, including therein the representative of the Pseudolioceras genus with variable morphologies, which considerably impairs the characteristic features of this species.

The holotype S. Buckman suggested was a small specimen from the collection of Licetta (1930; Plate 41 A, Figs. 1, 2). Typical (average) forms of P. *paracompactile sp. nov.* differ from it, given comparable diameters, in slightly more convex sides and better pronounced lower parts of the ribs. From the *P. compactile* metatype (Buckman 1909 - 1930; Plate 41 B, Figs. 1, 2), the new species differs in slightly more swollen whorls and noticeably larger umbilicus. From the topotypical specimen of *P. compactile* shown by A.A. Dagis (Dagis, 1974; Plate 18, Fig. 1) and coming from the Grammoceras thouarsense Zone in Yorkshire, England, the species under consideration differs in its slightly smaller umbilicus and less backward-bending ribs in the upper half of the whorl.

P. paracompactile sp. nov. is also close to the early Aalenian *P. maclintocki* (Haughton) that Ye.D. Kalacheva and I.I. Sei referred to as the *Tugurites* subgenus (Sei and Kalacheva, 1986). Its holotype, coming from Prince Patrick Island in the Canadian Arctic, is represented by a sketch and a brief description, which is insufficient for a clear impression of *P. maclintocki*. Hence, for the purposes of comparison, we used a collection of ammonites from Canadian Arctic islands (King Mackenzie Island, etc.) included by G. Frebold into the range of P. maclintocki. Following I.I. Sei and E.D. Kalacheva (1980), we accept as typical *P. maclintocki* those Pseudolioceras species that have well pronounced ribs all over the whorl (Frebold, 1960; Plate VII, Figs. 1, 2, 4 - 7, but not Plate VII, Figs. 3, 6 - 8).

With such an interpretation of *P. maclintocki*, the species described here differs from it in less well pronounced and slightly wider spaced ribs.

Range. Kolyma River basin, the Berezovka River; the upper Toarcian, the *Pseudolioceras paracompactile* Zone.

Pseudolioceras (Pseudolioceras) *evolutum* Repin *sp. nov.* Plate III, Figs. 2, 3, 6.

The name is derived from Latin evolvere (*evolutus*) meaning to roll out.

Holotype. Specimen 143/837, Plate III, Fig.3. The upper Toarcian, *P. danilovi* and *P. paracompactyle* zones in the Northeast of Russia.

Material. Twenty-one well-preserved cores and a few dozen casts and fragments of cores.

Description. A flattened medium-size shell (*d* is up to 50 mm) shows rapidly upward-growing and involute whorls. All over the last semi-whorl, the height grows by half, and the involution is up to 80 percent. The lateral sides are barely convex, grading smoothly into the narrow, keel-like venter. In shells greater than 20 mm in diameter, in the peri-umbilical areas of the lateral sides, there is a narrow depression. At this stage, the transition to tilted, almost vertical walls of the umbilicus is angular. In the interior whorls, the umbilical walls are rounded and low. The venter is rounded; in mature shells it expands and flattens near the mouth. The sides are covered by slightly sickle-shaped ribs, less pronounced near the umbilicus.

Comparison and remarks. Amid the known Pseudolioceras representatives, the new species has a wider umbilicus, making it closer to such species as P. alienum A. Dagis (Dagis and Dagis 1967; p. 54, Plate I, Figs. 7, 8) and P. kedonense Repin (Polubotko and Repin 1966; p. 52, Plate III, Figs. 1 - 3). Nevertheless, P. evolutum sp. nov. differs from P. alienum in a smaller umbilicus (umbilicus to diameter ratio is 20% against 26% in P. alienum), weakly concave sides near the umbilicus and generally coarser ribs.

In comparison with P. kedonense, the described species is characterized by a smaller number of coarser but less pronounced ribs in the lower half, and a more flattened venter. In sculpture and venter, P. evolutum sp. nov. is similar to the holotype of P. rosenkrantzi A. Dagis (Dagis and Dagis, 1965; p. 23, Plate I, Fig. 1), but has a wider umbilicus.

Species of the Pseudolioceras genus are characterized by noticeable polymorphism, with the sexual dimorphism being particularly apparent. The Toarcian - Aalenian Pseudolioceras species (P. lythense, P. kedonense, P. rosenkrantzi, P. beyrichi, etc.) have been shown to contain dimorphic pairs. Therefore, the newly described "species", P. evolutum, may in fact prove to be a virtually indistinguishable evolutoconch (microconch) of the species P. danilovi sp. nov. and P. paracompactile sp. nov., which may then account for its presence in both zones.

Range. The Kolyma River basin, the Berezovka River; the upper Toarcian, the P. danilovi and P. paracompactile zones.

REFERENCES

Buckman, S.S., *Yorkshire Type Ammonites*, London, 1930.

Dagis, A.A., *Toarskie Ammonity (Hildoceratidae) Severa Sibiri* (Toarcian Ammonites (*Hildoceratidae*) of Northern Siberia), Nauka, Novosibirsk, 1974.

Dagis, A.A and Dagis, A.S., On Subdivision of Toarcian Strata in the Northeast of the Soviet Union into Zones, *Stratigrafiya i Paleontologiya Mezozoiskikh Otlozhenii Severa Sibiri*, Nauka, Moscow, 1965, pp. 15 - 26.

Dagis, A.A. and Dagis, A.S., Stratigraphy of Toarcian Strata in the Vilyui Syneclise, *Problemy Paleontologicheskogo*

Obosnovaniya Detal'noi Stratigrafii Mezozoya Sibiri i Dal'nego Vostoka, Nauka, Leningrad, 1967, pp. 41 - 60.

Dietl, G. and Etzold, A., The Aalenian at the Type Locality, *Stuttgarter Beitr. Naturk.*, 1977, Ser. B., no. 30, pp. 1 - 13.

Frebold, H., The Jurassic Faunas of the Canadian Arctic. Lower Jurassic and Lowermost Middle Jurassic Ammonites, *Geol. Surv. Canada*, 1960, vol. 59, p. 33.

Polubotko, I.V. and Repin Yu.S., Stratigraphy and Ammonites of the Toarcian Stage in the Central Part of the Omolon Median Mass, *Materialy po Geologii i Poleznym Iskopaemym Severo-Vostoka SSSR*, Magadan, 1966, vol. 19, pp. 30 - 35.

Polubotko, I.V. and Repin, Yu.S., Lower and Middle Jurassic of the Northeast of the Soviet Union, *Newsl. Stratigr.*, 1988, vol. 19, no. 1/2, pp. 1 - 17.

Polubotko, I.V. and Repin, Yu.S., *Biostratigrafiya Aalenskogo Yarusa Severo-Vostoka SSSR* (Biostratigraphy of the Aalenian Stage in the Northeastern USSR), Nauka, Novosibirsk, 1974.

Repin, Yu.S., Lower and Middle Jurassic Ammonites in the Northeastern Part of the Soviet Union, *Polevoi Atlas Yurskoi Fauny i Flory Severo-Vostoka SSSR*, Magadan, 1968.

Resheniya 2-go Mezhvedomstvennogo Stratigraficheskogo Soveshchaniya po Dokembriyu i Fanerozoyu Severo-Vostoka SSSR (Resolutions of the 2nd Interdepartmental Stratigraphic conference on the Precambrian and Phanerozoic of the Northeast of the Soviet Union), 1974 - 1975, Magadan, 1978.

Resheniya 3-go Mezhvedomstvennogo Regional'nogo Stratigraficheskogo Soveshchaniya po Mezozoyu i Kainozoyu Srednei Sibiri, (Resolutions of the 3rd Interdepartmental Regional Conference on the Mesozoic and Cenozoic of Middle Siberia), 1978, Novosibirsk, 1981.

Sei, I.I. and Kalacheva, E.D., *Biostratigrafiya Nizhne- i Srednеyurskikh Otlozhenii Dal'nego Vostoka* (Biostratigraphy of the Lower and Middle Jurassic Strata of the Far East), Nedra, Leningrad, 1980.

Sei, I.I. and Kalacheva, E.D., The Jurassic Ammonite Pseudolioceras (*Tugurites*) of the Bering Province, *Can. J. Earth Sci.*, 1986, vol. 23, pp. 1024 - 1045.

Stratigrafiya Yurskoi Sistemy Severa SSSR (Stratigraphy of the Jurassic System of the Northern Soviet Union), Nauka, Moscow, 1976.

Zony Yurskoi Sistemy v SSSR (Zones of the Jurassic System in the Soviet Union), Nauka, Leningrad, 1982.

Stratigraphy and Geological Correlation, Vol. 1, No. 1, 1993. Translated from Stratigrafiya.Geologicheskaya Korrelyatsiya, Vol. 20, No. 2, 1993, pp. 118 - 127.
Original Russian Edition Copyright © 1993 by Stratigrafiya. Geologicheskaya Korrelyatsiya, Vangengeim, Pevzner.
English Translation Copyright © 1993 by Interperiodica.

Correlation of the Miocene Mammalian Reference Localities with the Magnetochronologic Scale

E. A. Vangengeim and M. A. Pevzner

Geological Institute, Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 109017 Russia

Manuscript received January 25, 1992

Abstract – Faunal, paleomagnetic, and geological analyses resulted in establishing the succession of some fossil mammalian localities of the middle Sarmatian, Meotian, and Pontian in the northern Black Sea region, Transcaucasia, and Kazakhstan. Features shared by Western European faunal events at the lower and upper boundaries of the Vallesian and at the upper boundary of the Turolian were elucidated, as well as some distinctive features of Vallesian and Turolian faunas from the studied areas.

INTRODUCTION

Biostratigraphers presently devote a great deal of attention to the development of the continental Neogene mammalian time scale and to the correlation of the faunas from different paleozoogeographical regions. The aim of this paper is to analyze the available data that can be used for developing the regional biostratigraphic scale for the eastern and central parts of Northern Palearctic and for establishing criteria for its correlation with the West European continental time scale (Mein, 1990).

Because the faunas of distant areas are sufficiently different, we need to determine the isochronous levels in faunal evolution and to analyze the associated events in order to develop the correlational criteria.

The paper is generally based on the results of magnetostratigraphic research of some late Neogene mammalian reference localities in the northern Black Sea region (Fig. 1). Localities that are associated with marine deposits of the Eastern Paratethys are of particular interest. Some mammalian localities without the paleomagnetic characteristics, but with rather precisely established geological position within the regional stages of the Eastern Paratethys, were also used. The availability of a magnetostratigraphic scale for the marine Neogene of the Eastern Paratethys (Vangengeim *et al.*, 1988; Pevzner and Vangengeim, 1984) enabled us to correlate these localities with the magnetochronologic scale.

The described stratigraphic interval corresponds to the Vallesian and Turolian of Western Europe.

FOSSIL MAMMALIAN LOCALITIES OF THE VALLESIAN

The first occurrence of *Hipparion* in the old world indicates the lower boundary of the Vallesian, and K-Ar age determinations of this event range from 11 to 11.5 Ma. Petrikany, Gidigich, Golbochika, and Kutuzov near Kishinev (Fig. 2) are the most ancient East European localities with hipparion remains. They yield the archaic form from the *H. primigenium* group. Other taxa present include *Microstonyx antiquus* (Kaup.), *Thalassictis robusta* (Nordm.), and others (Lungu and Chemyrtan, 1986). The bones were found in biohermal limestone from the lower part of the middle Sarmatian (Bessarabian).

The localities of Buzhor 1, Lapushna, and Kalfa in Moldova (which are nearly as old) have yielded only Bessarabian marine and deltaic deposits. The bone-bearing deposits are normally magnetized and correlated with the second half of polarity Chron 11. The fauna of these localities yields *Hipparion sarmaticum* Lungu (gr. *primigenium*), *Machairodus lascarevi* Lungu (gr. *aphanistus*), *Progonomys cathalai* Shaub, and others (Pevzner *et al.*, 1987), as well as some Astaracian forms.

Therefore, we can assign the age of the first hipparion occurrence in Eastern Europe to the interval between 10.8 Ma (top of Chron 11) and 11.4 Ma (base of the middle Sarmatian).

The bone beds of Buzhor 2 and Varnitsa in Moldova are directly above the reversal between Chrons 11 and 10 in the reversed polarity interval [Pevzner *et al.*, 1987]. *Percrocuta gigantea* (Schlos.), *Moldoredunca*, and the second form of hipparion, *Hipparion aff. verae*, appeared here. The number of Astaracian forms decreased, but *Lagomeryx* and *Euprox* continued to occur.

All the localities mentioned above formed near the coastline during the middle Sarmatian regression, when the sea in the Galician bay retreated from the latitude of Rovno at least to the Buzhor–Varnitsa line. Later, a new transgression extended northward, and the Galician Bay returned to its former dimensions.

The localities of Gritsev and Klimentovichi in Ukraine are nearly the same ages as Buzhor 2 and Varnitsa. In Gritsev, the bones of vertebrates occur in karst

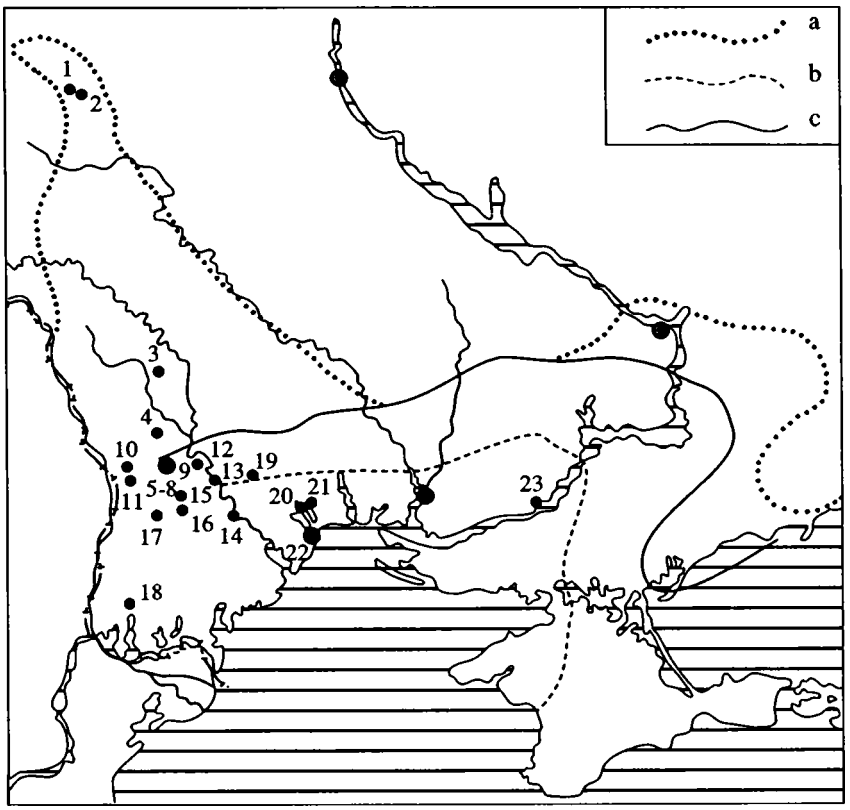


Fig. 1. *Mammalian localities in the northern Black Sea region.*
a – Presumed coastline in the second half of the middle Sarmatian, b – Presumed coastline at the beginning of the upper Sarmatian, c – Presumed coastline of the upper Sarmatian sea (according to V.P. Kolesnikov, 1940); Localities: 1 – Gritsev, 2 – Klimentovichi, 3 – Raspopeny, 4 – Poksheshty, 5 – Petrikany, 6 – Gidigich, 7 – Golbochika, 8 – Kutuzov, 9 – Atavaska, 10 – Buzhor, 11 – Lapushna, 12 – Kalfa, 13 – Varnitsa, 14 – Chobruchi, 15 – Kainary, 16 – Tarakliya, 17 – Chimishliya, 18 – Vinogradovka, 19 – Grebeniki, 20 – Novaya Emetovka, 21 – Staraya Kubanka, 22 – Odessa (16th Station of Bol'shoi Fontan, Pontian lectostratotype), 23 – Berislav.

fissures in biohermal limestones and in clays, which lie above the bioherms and contain the shells of middle Sarmatian molluscs. The bone-bearing deposits show a reversed polarity (Korotkevich, 1988). The fauna yields *Hipparion* ex gr. *primigenium, H.* sp., *Lagomeryx, Euprox, Amphilagus, Percrocuta (Dinocrocuta) gigantea, Dorcatherium*, and others (Korotkevich, 1988; Semenov, 1989; Lungu, personal communication, 1990).

Comparing the faunas of Klimentovichi and Gritsev, E.L. Korotkevich (1988, p. 110) stated that: "Common forms from both localities are very similar, if not identical, at the specific level. On these grounds, the Klimentovichi and Gritsev localities were assigned to the same faunal subdivision... and the age of these localities was determined as the beginning of the middle Sarmatian...." But the occurrence of two forms of hipparion in Gritsev and a reversed polarity of the enclosing deposits contradict this assignment. A paleomagnetic study of the complete middle Sarmatian sections showed that its lower and upper parts have a normal (Chron 11) and a reversed (Chron 10) polarity respectively. Therefore, the Gritsev and Klimentovichi localities cannot be older than the middle Bessarabian. The archaic character of their fauna (demonstrated by the occurrence of many forms common with the Astaracian) in comparison with the fauna of Varnitsa in Moldova, which is approximately of the same age, appears to result from their attribution to different landscape zones: the fauna of Varnitsa was dominated by representatives of open landscapes, while up to 80% of the fauna in Gritsev was inhabitants of warm humid forests (Korotkevich, 1988).

Among the late Sarmatian localities associated with exposed marine deposits, a paleomagnetic record was available for the localities El'dari in Transcaucasia and Kainary in Moldova.

The El'dari bone bed, according to its stratigraphic position, refers to the upper part of the Khersonian and is correlated with the upper part of Chron 9 (Vangengeim *et al.*, 1989). The El'dari fauna contains two forms of hipparion, i.e., *Hipparion eldaricum* Gab. (the group of *H. primigenium*) and *H.* sp. with a long protocone. It also contains *Diceros gabuniai* Zisk. (which are similar to the small archaic Vallesian rhinoceroses of Turkey and which were included in the *D. neumayri* species), *Chilotherium (Acerorhinus) zernovi* (Boriss.), *Percrocuta (Dinocrocuta) gigantea* (Schlos.), and oth-

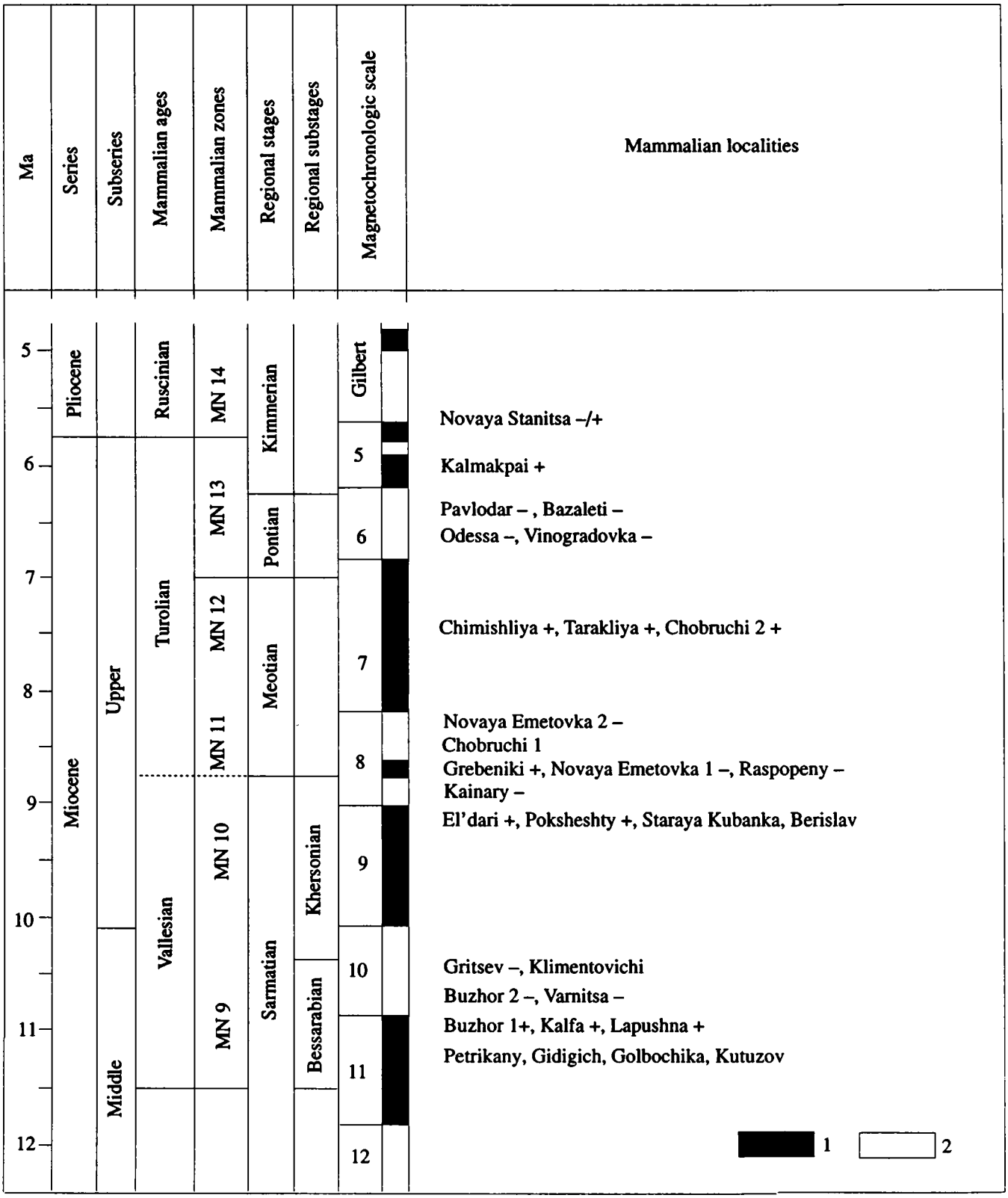


Fig. 2. *The position of mammalian localities in the magnetochronologic and stratigraphic scales.*
1 – normal polarity, 2 – reversed polarity; normal magnetization (+); reversed magnetization (–).

ers. The appearance of the archaic gazelles. *Graecoryx*, and *Procapreolus* is noteworthy. The giraffes are represented by *Palaeotragus (Achtiaria) borissiaki* (Alex.), one of the latest and most specialized forms of *Achtiaria* subgenus (Godina, 1979).

The locality Staraya Kubanka on the left bank of the Kuyal'nitskii Liman is also associated with the upper part of the upper Sarmatian marine deposits (Korotkevich, 1988). The locality Berislav in Ukraine, attributed by E.L. Korotkevich (1988) and Yu.A. Semenov (1989) to the beginning of the Khersonian, corresponds to the same stratigraphic level. Today, it is inaccessible because it is located within the Kakhovka Reservoir. The bone bed is associated with the upper Sarmatian marine deposits. Paleogeographical analysis suggests that this area was covered by the sea at the beginning of the upper Sarmatian. The coastline of the basin was located considerably farther to the north before the beginning of the late Sarmatian transgression (Fig. 1): therefore, the locality could have formed only during the regressive stage of the late Sarmatian basin, i.e., at the end of the Khersonian.

The faunas of these localities are rather similar. *Gazella, Procapra, Graecoryx*, and *Samotherium* appeared here. Hipparions are represented in Berislav by *Hipparion* cf. *verae* Gab. and in Staraya Kubanka by *H. verae* and *H. giganteum* Grom. These species are especially characteristic for the faunas of the terminal late Sarmatian and, most possibly, for the beginning of the Meotian.

The Poksheshty locality (Moldova) is confined to the continental deposits of the Balta Formation that lie over the marine middle Sarmatian. The bone bed is normally magnetized and correlated with Chron 9.

It is not possible to more precisely define the position of this locality within Chron 9 using only the paleomagnetic record. However, judging by the similarity between the faunas of Poksheshty and of Berislav, both localities appear to correspond to the same stratigraphic level, i.e., the top of the upper Sarmatian (the end of Chron 9).

We assign the locality Kainary in Moldova to the very end of the upper Sarmatian. It is confined to the reversely magnetized marine deposits that contain upper Sarmatian mollusk shells. Its fauna consists of *Progonomys woelferi* Bachm. et Wilson, *Parapodemus lugdunensis* (Schaub.), and others (Lungu, 1990). Using only the paleomagnetic record, we assign this locality either to the beginning or to the end of the Khersonian (the end of Chron 10 and the beginning of Chron 8 respectively) because only these upper Sarmatian intervals exhibit reversed polarity. However, because the section is regressive, we conclude that this locality was formed at the end of the Khersonian. It corresponds to the interval between the top of Chron 9 and the normal polarity event at the beginning of Chron 8.

The Kainary fauna has many forms in common with the fauna of the locality Kohfidisch in Austria, such as *Anurosorex kormosi* Bachm. et Wils., *Petenyia dubia* Bachm. et Wils., *Peteniella repenningi* Bachm. et Wils., *Spermophilinus bredai* (Mayr.), *Parapodemus lugdunensis* Schaub., *Progonomys woelferi* Bachm. et Wils., and *Kowalskia orientalis* Lungu (*K.* cf. *falbuschi*, according to P. Mein's identification), Kohfidisch corresponds to the very end of the Vallesian (Mein, 1990). Taking this resemblance into account, we can refer the Kainary fauna to the end of the Vallesian as well. In that case, the upper boundary of the Vallesian nearly coincides with that of the Khersonian (8.8 - 8.7 Ma).

Western European researchers place the top of the Vallesian in the magnetochronologic scale near the boundary of Chrons 9 and 8. Data on the Kastellios locality (Greece) places this boundary at the beginning of Chron 8 (Vangengeim *et al.* 1988). According to K-Ar dates from the Maraga locality (Iran), this boundary lies in the interval between 9.1 and 8.7 Ma (Bernor *et al.*, 1980).

FAUNAL CHARACTERISTICS OF THE VALLESIAN

A comparison of the Eastern European Vallesian faunas with those from Western Europe reveals the following resemblances and distinctions. The lower Vallesian faunas (zone MN 9) are generally quite similar. They include many forms that are characteristic of the earlier Astaracian faunal stage (*Lagomeryx* being especially abundant). Hipparions were represented by *Hipparion primigenium*; though by the end of this stage, their ecological differentiation occurred, and some localities yielded two forms of *Hipparion*. *Microstonyx* appeared at the very beginning of the Vallesian.

As in Western Europe, the Astaracian relics became quite rare in the upper Vallesian MN 10 zone; *Adcrocuta* appeared, as did *Parapodemus* at the very end of this stage.

In contrast to Western Europe, the earlier (end of the Vallesian) appearance of *Procapreolus, Gazella, Procapra, Graecoryx*, and *Microstonyx major* can be noted. These are open country forms, representatives of the typical Pikermi hipparion fauna. *Progonomys* may have appeared somewhat earlier than in Western Europe.

A large number of chilotheres, which often reached a subdominant level, was extremely characteristic of the Vallesian faunas of Eastern Europe, especially of late ones. The *Achtiaria* subgenus is also an important Vallesian taxon, and was later replaced by the *Palaeotragus* subgenus. In the second half of the upper Sarmatian, faunas developed a characteristic association of two hipparions *Hipparion verae* and *H. giganteum*, most likely surviving into the lower Meotian.

Presently, we cannot determine the position of the lower/upper Vallesian boundary (between MN 9 and MN 10 zones) with respect to regional stages of the Eastern Paratethys because there are no accurately dated mammalian localities in the interval from the top

of the Bessarabian to at least the middle of the Khersonian. In Western Europe, no dates have been determined for this boundary.

The appearance of some forms that are characteristic of advanced hipparion faunas in Sarmatian assemblages and the erroneous assigning of such localities as Berislav to the beginning of the upper Sarmatian have led some researchers (such as Korotkevich (1988) to correlate the upper Sarmatian with the lower Turolian (MN 11). A more accurate definition of the Berislav's age indicates that these forms appear only at the end of the upper Sarmatian, though a little earlier than in Western Europe. This fact indicates that the alternation of ecological groups was asynchronous in different regions (landscape zones) and, thus, cannot serve as a correlational criterion for the faunas of different provinces. We should note that some forms, characteristic of advanced hipparion faunas, were recorded in the lower Vallesian faunas from Turkey and Central Asia (where conditions were more arid than they were at that time in Europe) and in the regions adjacent to the Paratethys; in these localities; the main body of typical hipparion fauna formed earlier.

FOSSIL MAMMALIAN LOCALITIES OF THE TUROLIAN

Localities with fossil land mammals have not been recorded in the exposed Meotian marine deposits. However, in the northern Black Sea region some localities are associated with the continental strata, which are Meotian according to their stratigraphic position. They overlie the marine upper Sarmatian and are covered by lower Pontian marine deposits. We should note that the basal layers of this sequence can be referred to the end of the Sarmatian, when marine basin still occurred in southern regions and continental deposits accumulated in this area.

The Novaya Emetovka 1 locality (left shore of the Khadzhibei Liman) is one of the most ancient localities associated with this sequence. The bone bed is in the lower part of the alluvial-plain sequence. Its fauna yields hipparion with the morphological peculiarity of teeth, which approximate it to *Hipparion moldavicum* Grom. T.V. Krakhmal'naya (1987) placed it at the base of *H. moldavicum* Grom. lineage. *H. verae* Gab. and the first occurrence of *Palaeoryx* sp. (Krakhmal'naya, 1989) were recorded in this assemblage.

The bone-bearing deposits in Novaya Emetovka 1 are reversely magnetized (Krakhmal'naya *et al.*, 1988) and can be correlated with the beginning of Chron 8. Their precise position with respect to the normal polarity event at the beginning of Chron 8 cannot be defined, as it was not found in this section. If the bone bed is stratigraphically lower than this event, it should be dated as the very end of the Sarmatian; if higher, as the beginning of the Meotian because the Sarmatian/Meotian boundary coincides with the base of this event (Vangengeim *et al.*, 1988).

According to the systematic composition and morphological features of certain species, the mammalian assemblage of Novaya Emetovka 1 is transitional between the Turolian and Vallesian faunas (for example, the presence of *Hipparion* sp., which resembles both *H. moldavicum*, and *Achtiaria*).

In the Grebeniki locality (Odessa Oblast') the bone beds are confined to the lower part of continental deposits that overlie the marine upper Sarmatian. The faunal assemblage includes *Hipparion verae* Gab., *H. giganteum* Grom. (Gabuniya, 1959), *Machairodus ex gr. giganteum* (Sotnikova, personal communication 1990), and *Palaeotragus pavlowae* God. The latter, according to A.Ya. Godina (1979), combines features of *Palaeotragus* and *Achtiaria* in its skeleton. The age of this locality was defined differently: it was referred to the Meotian (Gabuniya, 1959) or to the upper Sarmatian (Korotkevich, 1988).

The bone-bearing deposits are normally magnetized and most likely suggest correlation with the normal event at the beginning of Chron 8. If this is true, these deposits should be assigned to the base of the Meotian.

Other geological position characterizes the locality Raspopeny (northern Moldova), which has rather similar fauna (Pevzner *et al.*, 1987). Bone beds occur within the terrestrial Balta Formation that overlies the middle Sarmatian marine deposits in this region. The mammalian assemblage is marked by the presence of highly specialized *Palaeotragus moldavicus* God., which is one of the latest representatives of the *Achtiaria* subgenus (Godina, 1979). Hipparions are represented by *H. giganteum*.

Deposits containing the bone remnants exhibit a reversed polarity formed at the beginning of Chron 8. The position of the bone bed with respect to the normal polarity event of Chron 8 was not defined for this locality or for Novaya Emetovka.

Most likely, the detailed morphological study of mammals, common for the three localities discussed above would enable us (according to the degree of species evolutionary development) to more precisely define the position of these localities with respect to one another and (in the magnetochronologic scale) with respect to normal polarity event of Chron 8, i.e., to the Sarmatian/Meotian boundary.

The Chobruchi locality (southeastern Moldova) contains two bone beds. The lowest, Chobruchi 1, is situated in the sandy granule member of the basal part of alluvial-plain sediments. The latter overlies the normally magnetized upper Sarmatian marine deposits and, in the lower reaches of the Dnestr River, lies under the marine Pontian. The paleomagnetic record of bone-bearing granule gravelstone from the lower horizon is not available. The gravelstone yields numerous fossil small mammals. The most important are *Valerimys sp.* and *Occitanomys neutrum* (Lungu 1990), According to L.K. Gabuniya (1959), *Hipparion* cf. *verae* and *Helladotherium duvernoyi* Gaud. seem to belong to the same bed. *Ictitherium* from the lower (?) horizon was

referred by Yu.A Semenov (1989) to *Ictitherium (Paraictitherium) pannonicum* Kretz.

The presence of *Valerimys, Occitanomys, Helladotherium*, and *Ictitherium pannonicum* indicates that the fauna can be referred to the Turolian.

The precise position of Chobruchi 1 with respect to the Sarmatian/Meotian boundary (as in the cases of the localities discussed above) cannot be established using the available data. However, the more progressive character of the Chobruchi fauna in comparison to that of the upper Sarmatian Kainary and their clear succession indicate that Chobruchi 1 can be referred with some certainty to the beginning of the Meotian.

The data reaffirm that the Vallesian/Turolian boundary approximates the Sarmatian/Meotian boundary.

The locality Novaya Emetovka 2 characterizes the next stratigraphic level. The bone bed from this locality is confined to the middle part of the alluvial sequence and lies 9 - 10 m above the lower bed, Novaya Emetovka 1. According to paleomagnetic record (Krakhmal'naya *et al.*, 1988), the upper bone-bearing horizon, Novaya Emetovka 2, corresponds to the same reversed polarity interval as the lower bed (Novaya Emetovka 1). We correlate this entire interval with Chron 8. Judging by the position in the section and in the magnetochronologic scale, this part of alluvial sequence can undoubtedly be referred to the first half of the Meotian.

The fauna of Novaya Emetovka 2 is characterized by *Adcrocuta eximia* (Roth. et Wagn.), *Ictitherium pannonicum* Kretz., *Hipparion moldavicum* Grom., *Palaeotragus rouenii* Gaud., and by the first appearance of *Procobus melania* Khom. (Korotkevich, 1988; Krakhmal'naya, 1989; Semenov, 1989). It is a typical Turolian fauna, as indicated by the presence of all these species and by the general aspect of the fauna.

The well-known localities Tarakliya and Chimishliya in Moldova are associated with the upper part of the pre-Pontian continental sequence. The bone-bearing deposits are normally magnetized and were formed during Chron 7, which we correlate with the upper part of the Meotian. The available data does not enable us to precisely define the position of these bone beds within Chron 7. Unfortunately, the fauna from these localities (with the exception of some groups) has not yet been studied sufficiently and the previous identification need to be revised. Both localities yield abundant antelopes, including Tragocerini, *Hyaenictitherium venator* Sem., *Ictitherium pannonicum* Kretz., *Hipparion moldavicum* Grom., and *Palaeotragus rouenii* Gaud. Moreover, *Samotherium* have been found in Tarakliya (Gabuniya, 1959; Semenov, 1989). This assemblage is typical of the advanced Turolian. The normally magnetized upper bone bed from the locality Chobruchi 2, which yields *Hipparion moldavicum*, corresponds to the same level.

Fossil mammalian localities associated with Pontian marine deposits had been found in Odessa (the 16th station of Bol'shoi Fontan, the Pontian lectostratotype) and near Bolgrad (Vinogradovka) in Ukraine. Their bone beds are situated in the lowest Pontian horizons and exhibit a reversed polarity formed during Chron 6. The fauna consists entirely of small mammals (Topachevskii *et al.*, 1988). The Odessa fauna closely resembles that of the Hungarian locality Polgardi, which is referred to the late Turolian. The Vinogradovka fauna appears to include more xerophilous biotopes. These faunas are marked by the appearance of *Polonomys*, a genus hitherto unknown from localities preceding the Ruscinian.

The large mammals in the Pontian marine deposits are represented by a few finds in the northern Black Sea region (Gabuniya, 1959; Korotkevich, 1988). They include *Deinotherium giganteum* Kaup., *Aceratherium incisivum* Kaup., *Diceros pachygnathus* Wagn., *Paracamelus* sp. and others.

Two localities with paleomagnetic records, Bazaleti in Georgia (Vangengeim *et al.*, 1989) and Pavlodar (Gusinyi Perelet) in Kazakhstan (Vangengeim *et al.*, 1984), are situated in the Pontian continental counterparts. The bone beds in both localities exhibit a reversed polarity.

Some endemic forms of cavicorns, together with typical northern Eurasian Turolian genera are present in Bazaleti (Meladze, 1967).

The well-known Pavlodar assemblage is an Asian variety of late hipparion fauna. In addition to genera that are widely distributed in Eurasia (such as *Adcrocuta, Chilotherium, Plesiogulo*, and others), it includes many forms in common with the Turolian dorcadoides fauna of Northern China (Kurten, 1952) (such as *Yuorlovia asiaticus* God., *Martes poleosinensis, Sinotherium, Hyaenictitherium*, and a rather endemic assemblage of rodents).

In the magnetochronologic scale, the Odessa and Vinogradovka localities can be confidently placed in the beginning of Chron 6, whereas they are reliably correlated with the Eastern Paratethys marine stratigraphic scale. At present, the precise position of Pavlodar and Bazaleti within Chron 6 cannot be defined.

The locality on the Kalmakpai River (Eastern Kazakhstan), confined to the Karabulak Formation, is stratigraphically higher. Its fauna includes late representatives of *Adcrocuta, Hyaenictitherium, Plesiogulo*, and others. The evolutionary level of forms common with the Pavlodar fauna also confirms the post-Pavlodar age of the Kalmakpai locality. Evidently, this fauna can be referred to the end of the Turolian. The bone beds are normally magnetized (Yakheemovich *et al.*, 1990) and correlate with the beginning of Chron 5.

The Novaya Stanitsa locality on the Irtysh River is assigned to the end of Chron 5 and the beginning of the Gilbert epoch. The fauna yields *Baranomys, Microtodon*, and *Promimomys*, which are genera characteristic of the Ruscinian (Zykin *et al.*, 1991).

The K-Ar dates are not presently available for the Turolian upper boundary. Various authors place it in the magnetochronologic scale near the upper boundary of Chron 5. The data on the last two localities indicate that

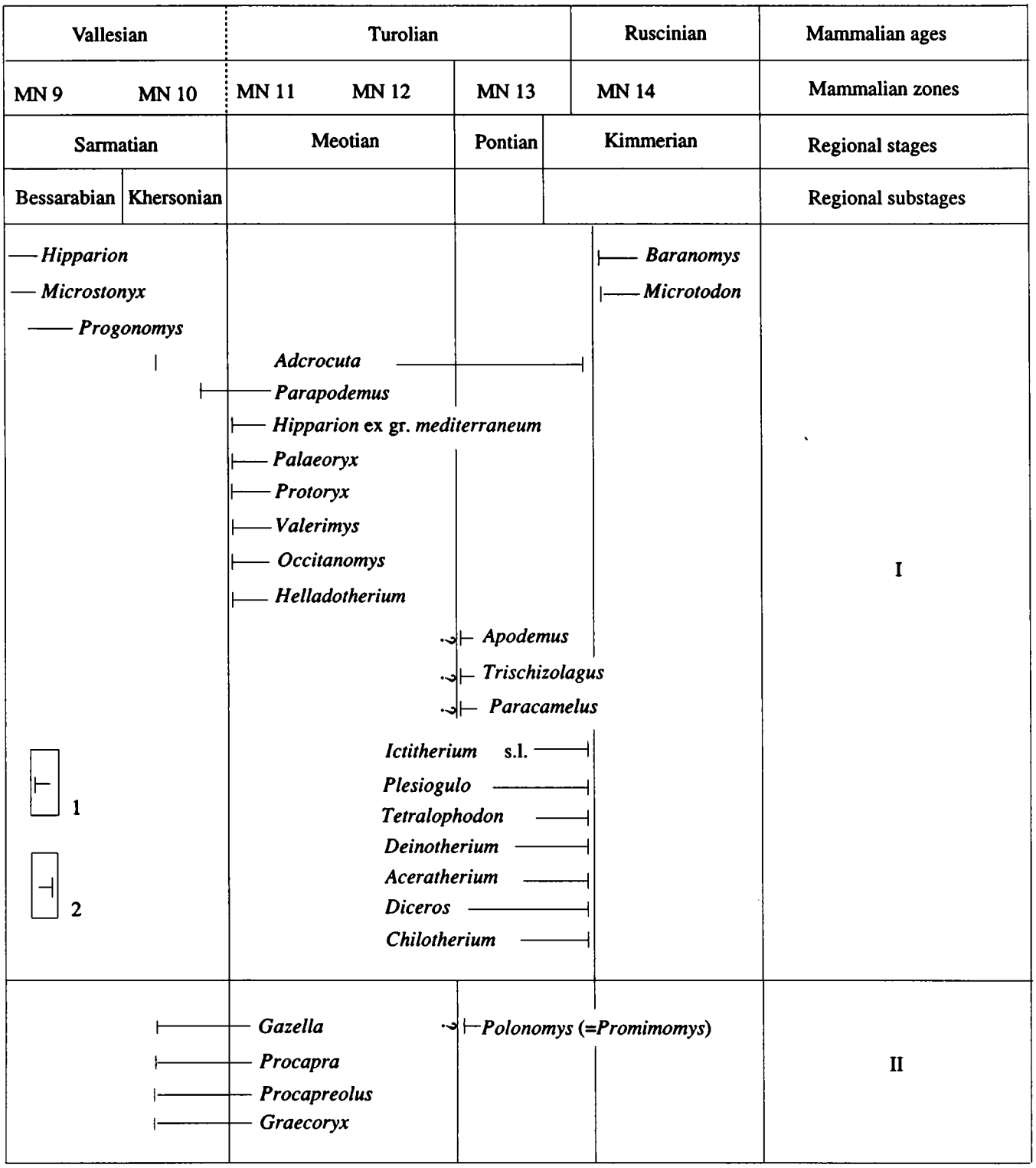


Fig. 3. *The appearance and extinction of some characteristic mammals in the eastern and central parts of northern Palearctic. I – simultaneously with Western Europe, II – appearance earlier than in Western Europe, 1 – first occurrence, 2 – extinction.*

the upper Turolian boundary occurs within Chron 5, and its age is approximately 5.5 - 5.6 Ma.

FAUNAL CHARACTERISTICS OF THE TUROLIAN

The Turolian fauna of Eastern Europe and Northern Asia was generally similar to Western European assemblages from the relatively arid zone. As in Western Europe, the Turolian was a flourishing period for the typical inhabitants of the savanna and savanna-steppe biotopes. Its beginning was marked by the appearance of *Valerimys, Occitanomys, Helladotherium, Protoryx,* and *Hipparion* ex gr. *mediterraneum.* Also noteworthy was the presence of *Microstonyx major* (Gerv.), numerous gazelles, Tragocerini, Hippotragini, and the wide distribution of Palaeotraginae. The number of chilotheres decreased in comparison with the Vallesian faunas: they yielded to the aceratheres, though they remained a typical member of assemblages in the Asian part of the Palearctic up to the end of the Turolian, sometimes reaching a subdominant level (for example, in Kalmakpai).

As in Western Europe, some genera (i.e., *Tetralophodon, Deinotherium, Chilotherium, Aceratherium, Plesiogulo, Adcrocuta,* as well as all ictitheres and some others) were extinct at the Turolian/Ruscinian boundary.

In contrast to Western Europe, where hipparions were extremely diverse (up to 5 species), not more than 3 species occurred simultaneously in Eastern Europe. *Samotherium* was extinct in Western Europe at the beginning of the Turolian, while it survived up to the terminal Turolian in Eastern Europe and Kazakhstan.

In the Western European stratigraphic scale, the Turolian is subdivided into three zones: MN 11-13. Our data enables us to establish zone MN 13 with the most certainty because of the appearance of "*Cricetus,*" *Trischizolagus, Apodemus,* and *Paracamelus.* Its lower boundary is not younger than that of the Pontian regional stage of the Eastern Paratethys and the base of Chron 6 (6.9 Ma).

As regards the separation of the MN 11 and MN 12 zones, we do not yet have enough data to solve this problem: there are no definite faunal criteria for this boundary and no radiometric dates. Nevertheless, we can observe in the Eastern European faunas two groups of localities with somewhat different faunas. The most ancient group (Grebeniki, Novaya Emetovka 1, Chobruchi 1, and, possibly, Raspopeny) has transitional Vallesian/Turolian fauna with some typical Turolian forms. The second group (Novaya Emetovka 2, Tarakliya, and Chimishliya) is characterized by advanced Turolian fauna. The boundary between these faunal groups is somewhat older than the base of Chron 7 (more than 8 Ma). Future research will show whether this boundary corresponds to the boundary between the MN 11 and MN 12 zones.

CONCLUSION

An analysis of the faunal and paleomagnetic data on the mammalian reference localities from Eastern Europe and some regions of North Asia, as well as the comparison with Western European record showed that the important faunal limits, boundaries of the stages of the continental scale, were isochronous. The appearance of migrants, which were not directly connected with drastic environmental changes, and the evolutionary development of common forms should serve as criteria for the correlation of Eastern and Western European faunas. The events shared by both Western and Eastern Europe are: (1) the appearance of *Hipparion* and *Microstonyx* – base of the Vallesian; (2) the appearance of *Valerimys, Occitanomys, Helladotherium,* and *Hipparion* ex gr. *mediterraneum* – base of the Turolian (in Eastern Europe, the evolutionary transitions *Palaeotragus (Achtiaria)* → *P. (Palaeotragus)* and *Machairodus* ex gr. *aphanistus* → *M. giganteus* occurred at the same boundary); (3) the extinction of *Deinotherium, Tetralophodon, Chilotherium, Aceratherium, Plesiogulo, Ictitherium* s.l., and *Adcrocuta* and the appearance of *Baranomys* and *Microtodon* – the top of the Turolian (Fig. 3).

REFERENCES

Bernor, R.L., Woodburne, M.O., and Van Couvering, J.A., A Contribution to the Chronology of Some Old World Miocene Faunas Based on Hipparionine Horses, *Geobios.*, 1980, no. 13, fasc. 5, pp. 705 - 739.

Gabuniya, L.K., *K Istorii Gipparionov* (On the History of Hipparions), Izd. AN SSSR, Moscow, 1959.

Godina, A.Ya., *Istoricheskoe Razvitie Zhiraf. Rod Palaeotragus* (The Historical Development of the Giraffes. Genus *Palaeotragus*), Nauka, Moscow, 1979.

Kolesnikov, V.P., *Verkhnii Miotsen* (Upper Miocene), Stratigrafiya SSSR, vol. 12, Neogen SSSR, Izd. AN SSSR, Moscow, Leningrad, 1940, pp. 229 - 373.

Korotkevich, E.L., *Istoriya Formirovaniya Gipparionovoi Fauny Vostochnoi Evropy* (History of Formation of the East European Hipparion Fauna), Naukova Dumka, Kiev, 1988.

Krakhmal'naya, T.V., Hipparions of the Late Miocene Fauna from Novaya Emetovka, *Materialy po Nekotorym Gruppam Pozdnekainozoiskikh Pozvonochnykh Ukrainy,* Kiev, 1987, (Preprint AN USSR, Zoological Inst. 87.12).

Krakhmal'naya, T.V., *Kopytnye Pozdnemiotsenovykh Terriofaun Novoi Emetovki* (Hoofed Theriofaunas of Novaya Emetovka in the Late Miocene), Cand. Sci. (Biol.) Dissertation, Zoological Inst. Ukrainian Academy of Sci., Kiev, 1989.

Krakhmal'naya, T.V., Svetlitskaya, T.V., and Chepalyga, A.L., New Data on the Theriofauna and Geology of Novaya Emetovka, *Vestnik Zoologii,* 1988, no. 5, pp. 32 - 37.

Kurten, B., The Chinese Hipparion Fauna, *Comm. Biol. Soc. Sci. Fenn.*, 1952, vol. 13, no. 4.

Lungu, A.N., *Rannie Etapy Razvitiya Gipparionovoi Fauny Kontinental'nogo Obramleniya Paratetisa* (Early Stages of the Hipparion Fauna Evolution in the Paratethys Continental Margins), Doctorate (Geol.-Min.) Dissertation, Geological Inst., Georgian Academy of Sci., Tbilisi, 1990.

Lungu, A.N. and Chemyrtan, G.D., Land Vertebrate Fauna of the Reef Stripe in the Kishinev Vicinity, *Pliotsen-Antropogenovaya Fauna Dnestrovsko-Prutskogo Mezhdurech'ya*, Shtiintsa, Kishinev, 1986, pp. 42 - 50.

Mein, P., Updating of MN Zones, *European Neogene Mammal Chronology*, Lindsay, E.H. *et al.*, Eds., Plenum Press, New York, 1990, pp. 73 - 90.

Meladze, G.K., *Gipparionovaya Fauna Arkneti i Bazaleti* (Hipparion Fauna of Arkneti and Bazaleti), Metsniereba, Tbilisi, 1967.

Pevzner, M.A., Lungu, A.N., Vangengeim, E.A., and Basilyan, A.E., Position in the Magnetochronologic Scale of the Vallesian Hipparion Localities in Moldova, *Izv. AN SSSR, Ser. Geol.*, 1987, no. 4, pp. 50 - 59.

Pevzner, M.A. and Vangengeim, E.A., Correlation of the Late Miocene West European Continental Time Scale with Stratigraphic Scales of the Mediterranean and Eastern Paratethys, *Izv. AN SSSR, Ser. Geol.*, 1984, no. 5, pp. 69 - 79.

Semenov, Yu.A., *Iktiterii i Morfologicheski Skhodnye Gieny Neogena SSSR* (Ictitheres and Morphologically Similar Hyenas in the USSR Neogene), Naukova Dumka, Kiev, 1989.

Topachevskii, V.A., Chepalyga, A.L., Nesin V.A., *et al.*, Microtheriofauna (Insectivora, Lagomorpha, Rodentia) of the Pontian Lectostratotype, *Dopov. AN USSR*, 1988, no. 4, ser. 5, pp. 90 - 93.

Vangengeim, E.A., Gabuniya, L.K., Pevzner, M.A., *et al.*, Stratigraphical Position of Transcaucasian Localities of the Hipparion Fauna according to Magnetostratigraphic Data, *Izvestiya AN SSSR, Ser. Geol.*, 1989, no. 8, pp. 70 - 77.

Vangengeim, E.A., Muskhelishvili, L.V., Pevzner, M.A., *et al.*, Upper Boundary of the Vallesian and Its Position in the Stratigraphic Scale of the Eastern Paratethys, *Izv. AN SSSR, Ser. Geol.*, 1988, no. 6, pp. 67 - 74.

Vangengeim, E.A., Zazhigin, V.S., Pevzner, M.A., and Khorevina, O.V., Miocene/Pliocene Boundary in West Siberia and Central Asia, *Sreda i Zhizn' na Rubezhakh Epokh Kainozoya v Sibiri i na Dal'nem Vostoke*, Nauka, Novosibirsk, 1984, pp. 167 - 171.

Yakheemovich, V.L., Suleimanova, F.I., Borisov, B.A., *et al.*, Neogene Correlation of East Paratethys, Zaisan Depression, and Arctic Basin via Fore-Urals, *Global Events and Neogene Evolution of the Mediterranean*, Abstracts Inst. Paleont. Sabadell, 1990, pp. 375 - 376.

Zykin, V.S., Zazhigin, V.S., and Kazanskii, A.Yu., Late Neogene of the Southern West Siberian Plain: Stratigraphy, Paleomagnetism, Main Climatic Events, *Geologiya i Geofizika*, 1991, no. 1, pp. 78 - 86.

Stratigraphy and Geological Correlation, Vol. 1, No. 1, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 20, No. 2, 1993, pp. 128 - 136.
Original Russian Edition Copyright © 1993 by Stratigrafiya. Geologicheskaya Korrelyatsiya, Zaporozhets.
English Translation Copyright © 1993 by Interperiodica.

Lower Oligocene Palynology and Phytoplankton of the Northern Caucasus

N. I. Zaporozhets

Geological Institute, Russian Academy of Sciences, Pyzhevskii Per. 7, Moscow, 109017 Russia

Manuscript received March 5, 1991

Abstract – Characteristics have been given to the palynological assemblages and organic-walled phytoplankton of the Pshekha and Solenovskii horizons, and the basal strata of the Morozkina Balka Beds of the Kuban' reference section and strata of the same time interval, penetrated by hole 32 within the Stavropol' arch. They are compared to the corresponding assemblages of the lower Oligocene deposits of Armenia (Landzhar section) and Northern Ustyurt (Ashcheairyk and Chegan sections).

The Paleogene sequence at the left bank of the Kuban' River opposite the city of Cherkessk, is the most representative of the southern ones in the former USSR territory as it is the richest in organic remains, particularly microplankton. According to planktonic foraminifers and nannoplankton, its individual parts are consistent with the Paleogene subdivisions of the general chronological scale, which can be considered as a reference when correlating the deposits of various facies zones of the Crimea-Caucasian region and of adjacent Kazakhstan.

In the paper, the lower Oligocene sequence of the Kuban' section is subdivided based on dinocysts, green algae, pollen and spores[1]. The analyzed complexes are compared to the synchronous ones from the marine deposits of the Erevan-Ordubad zone of Armenia and of the Northern Ustyurt basin.

The analyzed strata include the upper part of the Belaya Glina Formation assigned to the upper Eocene according to planktonic foraminifers and nannoplankton, namely, *Globorotalia centralis – Globigerina gortanii* (P 17) and *Discoaster barbadiensis* (NP 18 - 20) zones, respectively. The Pshekha Horizon, Ostracoda Bed (Polba R. strata), and clay of the lowermost Morozkina Balka Beds (Fig. 1) are referred to as the lower Oligocene, as the base of the Pshekha Horizon contains the assemblage of planktonic foraminifers of the *Globigerina tapuriensis* (P 18) Zone and nannoplankton of the *Ericsonia subdisticha* and *Helcosphaera reticulata* (NP 21 and 22) zones. The Ostracoda Bed and overlying clay are assigned to the *Sphenolithus predistentus* (NP 23) Zone (Gotsadze and Chikhovani, 1989; Krasheninnikov and Muzylev, 1975). The total thickness of the analyzed section is about 50 m, 40 of which are taken up by the Pshekha Horizon, and 3 m belonging to the Ostracoda Bed.

Besides the Kuban' section, Samples were also taken from the lowermost Maikop group, directly at its contact with the Belaya Glina Formation, and also from the Ostracoda Bed, penetrated by hole S-29 in the southwestern Stavropol' arch, where they were taken from the depths of 242 and 205 m.

The Belaya Glina Formation in the Kuban' River valley is presented by dense or lumpy white, sometimes greenish-white marls, whereas the Pshekha Horizon are the alternation of light-grey or olive-grey calcareous or slightly calcareous clay, with ocher pigmentation coating due to jarosite. Scattered siderite nodules, abundant deformed planorbell tests, skeletal remains and gill fragments of fish may occur in the clay. The Ostracoda Bed is composed of light-grey non-laminated marl, formed mainly by the remains of coccolithophorides and Ostracoda. The overlying beds of the Morozkina Balka Beds are presented by dark-grey clay with thin interbeds of lighter marl.

The palynologic complex, Sampled one meter below the top of the Belaya Glina Formation (Sample 18-b) from the white marls with greenish pigmentation, is mainly presented by pine pollen (>95%) with small quantities of Taxodiaceae and rare pollen of *Cedrus*, *Glyptostrobus*, *Sciadopitys*, *Ephedra* (in particular *E. cheganica*), *Podocarpus*, *Quercus*, *Magnolia* and spores of Polypodiaceae.

The organic-walled phytoplankton is very rich: it is presented by 83 various taxons (4 of them are green algae). Quantitatively, phytoplankton exceeds the number of palynomorphs of high plants by 2.5 times and is considerably more diverse. A high marine coefficient suggests that sediments were formed in the offshore environment with the normal salinity of the open sea[2]. A complete reduction of the *Wetzeliella-Charlesdownia-Kisselevia* group specimens, including the index-species of the upper Eocene *Charlesdownia clathrata angulosa* Zone,

[1] The Maikop sequence of the Kuban' section is given according to V.A. Grossgeim (Stratigraphic Glossary for the USSR Territory, The Paleogene, Neogene, Quaternary, Leningrad, Nedra, 1982, pp. 271 - 272).

[2] The marine coefficient is the ratio of marine organic-walled phytoplankton quantity to the total number of palynomorphs (spores and pollen + organic-walled phytoplankton).

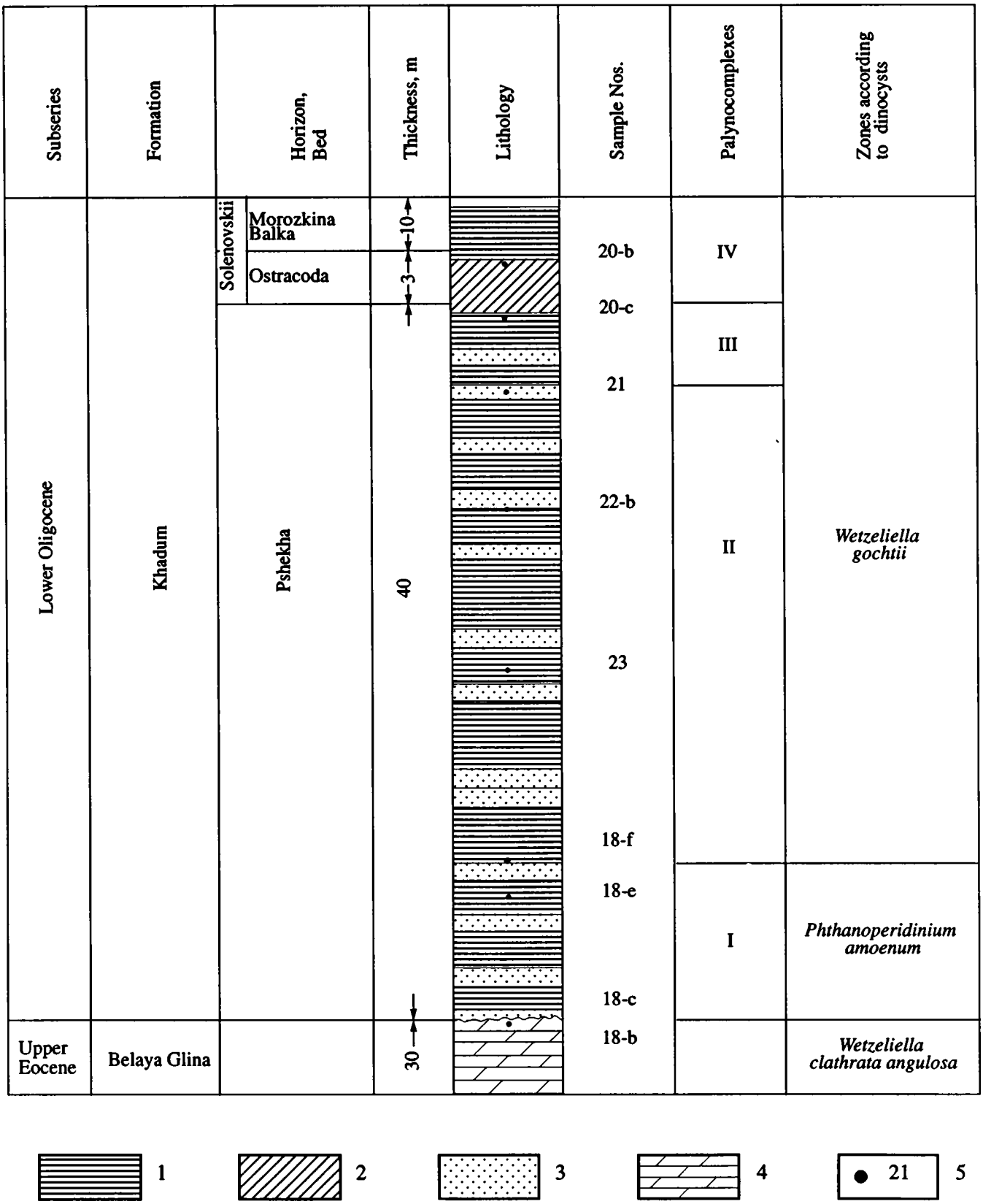


Fig. 1. *Lower Oligocene sequence on the left bank of the Kuban' River near the city of Cherkessk.*

1 – clay and silt, 2 – marl of Ostracoda Bed, 3 – sand and loosely cemented sandstone, 4 – carbonate clay, 5 – site of sampling and its number.

characterizes the above-mentioned complex and differentiates it from one of the Eocene uppermost horizons in Kazakhstan and Armenia. However, the presence of green algae *"Lanjaria"* and *Titthodiscus beloglinensis*, and also of other dinoflagellate species composing the zonal complex *Areosphaeridium diktyoplokus, A. arcuatum, Deflandrea leptodermata, Thalassiphora pelagica, Dapsilidinium pastielsii, Pyxidiella scabrosa, Lejeunecysta hyalina*, and others allows the dating of the uppermost Belaya Glina Formation beds as the late Eocene. The bulk of the complex is composed of *Impagidinium, Cleistosphaeridium, Pyxidiella*, cf. *Canningia, Spiniferites* (abundant, in particular, *S. ramosus* and *S. pseudofurcatus* and their subspecies), *Thalassophora, Deflandrea, Microdinium, Operculodinium, Pentadinium*, and *Tectatodinium*. Rare specimens of *Systematophora, Hystrichokolpoma* (*H. salacia* and *H. cinctum*), *Homotriblium, Selenophemphix* (*S. nephroides* and *S. selenoides*), *Phelodinium, Lejeunecysta, Lentinia, Lophocysta, Pentadinium laticinctum, Batiacasphaera, Lingulodinium, Heteraulacacysta campanula, Milliododinium tenuitabulatum, Planoperidinium gracile, Cordosphaeridium latispinosum, C.microtriainum, Eisenackia, Cannosphaeropsis, Fibrocysta, Oligosphaeridium*, and others are also present.

A Sample from hole S-29 taken directly from the base of the Pshekha Horizon characterizes the littoral environment where palynomorphs of the high plants predominate sharply among plant microfossils. The ratio of them to phytoplankton is 60 : 1. The pollen-spore complex is rich and has more than 60 taxons. The gymnosperms' pollen sharply dominates (>90%), however, the Taxodiaceae are as abundant as the pine ones. The pollen of that group has in its spectrum the species typical of the Oligocene, namely, *Tsuga crispa* and *Picea tobolica*. The angiosperms (>10%) show a predominating role of Betulaceae, especially hornbeam and Fagaceae pollen (*Quercus sp., Q. graciliformis, Castanopsis*), with a considerable share of *Yuglandaceae* and sumach. The rest of taxons, namely, *Acer, Liquidambar, Comptonia, Myrica*, Ericaceae, *Ilex*, etc. are scarce. In the composition of palynomorphs, the described complex resembles that of the lowermost Pshekha Horizon of the Kuban' section.

Phytoplankton is scanty and contains forms and species the majority of which are transient from the Eocene to Oligocene, for instance, *Lejeunecysta hyalina, Phelodinium. Deflandrea, Diconodinium, Palaecystodinium golzowense*, cf. *Scriniodinium*; among them green algae *Ptanospermella* and *Cymatiosphaera* are also present.

The palynomorph complex of high plants from the Pshekha Horizon of the Kuban' section reflects, in fact, the temperate-thermophilic flora with participation of subtropical forms typical of the early Oligocene from the southern parts of the Boreal region, where deciduous, coniferous broad-leaved mesophilous forests were widespread. Some arid phases, poorly expressed in the Northern Caucasus, can be recognized by a slightly higher occurrence of *Ephedra* pollen in combination with the abundant pine and sumach pollen in some intervals of the Kuban' section. At least three palynocomplexes may be identified there with respect to the proportion of various palynomorphs, dominating ones, in particular, which vary from one interval of the reference section to another; they characterize the basal, middle and upper parts of the Pshekha Horizon.

The lower palynocomplex is characterized by Samples 18-c and 18-e, taken from the very base of the Maikop Group and 7.0 m over its base. Compositionally, it resembles that Sampled from the depth of 242 m at hole S-29. The described interval is characterized by an obvious predominance of the Gymnospermae pollen that is typical of the Eocene/Oligocene boundary in the sections of the southern parts of the former Soviet Union. Within the Kuban' region, the Gymnospermae pollen composes from 80 to 92%, the pine pollen dominating at the base of the sequence; whereas Taxodiaceae prevails up-section. The Angiospermae spectrum has the dominating amount of Fagaceae and Hamamelidaceae pollen, and upsection, of sumach (up to 45%), oak, and Tricolporopollenites (20% of each in the total amount of Angiospermae pollen). Besides the listed above taxons, there occur small amounts of the pollen of *Cedrus, Sciadopitys* (up to 2%), *Glyptostrobus, Ephedra fusiformis, E. cheganica*, and also other species of *Ephedra* (1%), *Podocarpus*, in particular, *P. nagaeformis, Castanopsis, Fothergilla, Hamamelis* (in particular, *H. scotica*), *Corylopsis, Liquidambar, Carya, Juglans, Engelhardtia, Cyclocarya, Myrica, Nyssa, Magnolia, Palmae, Platanus, Araliaceae, Tilia, Ilex*, Moraceae. Sample 18-e shows considerable participation of Betulaceae pollen, particularly, of the hornbeam; there appears a pollen of *Carya glabraiformis, Juglans sieboldianiformis*, and others typical of the Oligocene. In the Juglandaceae spectrum, a more appreciable role is played by *Engelhardtia*.

The dinocyst assemblage of the analyzed interval is extremely diverse (over 80 species), though quantitatively it is not so abundant as the phytoplankton of the Belaya Glina Formation, taking up less than 10% of all the palynomorphs. On the whole, the assemblage is characterized by a combination of the species, transient from the Eocene and represented both in the early Oligocene and in the Oligocene as a whole, with the species that became extinct in its beginning. There are rare specimens of *Wetzeliella symmetrica*, and also the form that resembles the *W. gochtii*, that usually appears at higher stratigraphic levels of the Oligocene. Therefore, more detailed study of that taxon is required. The majority of the phytoplanktonic species is the open sea inhabitants, and only the abundant appearance of cf. *Canningia* suggests a certain variation of the sea regime in the sedimentation basin. The most abundant and variable taxons are *Deflandrea* (*D. phosphoritica, D. leptodermata, D. heterophlycta, D. arcuata*), *Areosphaeridium* (*A. diktyoplokus, A. pectiniforme, A. arcuatum*), *Spiniferites* (*S. pseudofurcatus, S. ramosus, S. wetzelii*). The rest of the species are represented by rare *Selenophempix nephroides, Samlandia chlamidophora, Pyxidiella scabrosa, Palaeocystodinium golzowense, Len-*

tinia cf. wetzelii, Thalassiphora delicata, Th. pelagica, Lingulodinium, Operculodinium, Cleistosphaeridium, Hystrichokolpoma salacia, H. cinctum, H. rigaudiae, Milliododinium tenuitabulatum, Dapsilidinium pastielsii, Tectatodinium, Cannosphaeropsis, Systematophora, Cordosphaeridium microtriatum, Selenophemphix coronata, S. nephroides, Diphyes colligerum, Impagidinium, Microdinium, Turbiosphaera, Stoverocysta, Rottnestia borussica, Homotriblium sp., *H. floripes, Lejeunecysta globosa*, and others (Plate I). That part of the section is attributed to the *Phthanoperidinium amoenum* Zone. Though the index species proper were met as rare specimens only in Sample 18-e, the rest of the assemblage composition of that interval is evidence that it belongs to that zone. In the group of green algae and acritarchs, there is a scarce occurrence of *Crassosphaera, Tytthodiscus beloglinensis*, and *Micrhystridium*. The boundary between the *Phthanoperidinium amoenum* and *Wetzeliella gochtii* Zones is situated within a 2-m interval of the section, restricted by Samples 18-e and 18-f, coinciding provisionally with the boundaries of palynocomplexes I and II. *Tytthodiscus beloglinensis*, which is an extremely rare taxon in the Oligocene, became extinct at that level. The specific variety of *Phthanoperidinium* and *Heterolaucacysta* is reduced there. As for the phytoplankton composition in Sample 18-f, it does not seem to essentially differ from that characteristic of the bulk of the Pshekha Horizon overlying the bed marked by Sample 18-e.

The palynocomplex of the middle part of the Pshekha Horizon (Samples 18-f, 19, 22, and 23) is characterized by a distinct predominance of *Gymnospermae* pollen, in particular, Taxodiaceae, and by its reduction in the highest of two levels. In the *Angiospermae* spectrum, the pollen of Fagaceae, sumach and two types of the *Tricolporopollenites* dominates. In contrast to the complexes of other intervals of the Pshekha Horizon, the role of the thermophilic elements is extremely small: there occur only *Castanopsis, Quercus graciliformis, Q. Conferta, Engelhardtia, Moraceae, Liquidambar, Hamamelis*, and *Araliaceae*. Taxons known from the lower Oligocene deposits in the southern regions of the former Soviet Union, Kazakhstan, and Western Siberia, were scarce, but recognized in the considered palynocomplex as well as in the Sample taken from the depth of 242 m at hole S-29. They are: *Tsuga crispa, T. tadaeformis, Ephedra cheganica, Quercus williamsoniana, Juglans sieboldianiformis*, and *Carya glabraeformis*.

Phytoplankton of the middle part of the Pshekha Horizon is scarce but diverse (about 40 species). No dominating taxons have been recognized there. Both index species of the early Oligocene occur there, namely, *Phthanoperidinium amoenum* and *Wetzeliella gochtii*, and also *W. symmetrica*. Thus, the described interval can be undoubtedly included in the *W. gochtii* Zone. The presence, though scarce, of *Pediastrum* and *Micrhystridium* in combination with *Azolla*, may indirectly indicate the slight freshening of the basin. On the whole, the complex includes the following taxons: *Spiniferites ramosus, Lejeunecysta paratenella, L. hyalina, Hystrichostrogilon membraniforum, Batiacasphaera baculata, Areosphaeridium pectiniforme, A. arcuatum, Heteraulacacysta campanula, Palaeocystodinium golzowense, Dapsilidinium pastielsii, Hystrichokolpoma rigaudiae, Cordosphaeridium inodes, Phthanoperidinium comatum, Deflandrea phosphoritica, Thalassiphora pelagica, Tectatodinium, Achomosphaera*, cf. *Cannosphaeropsis, Milliododinium, Homotriblium*, and others.

Palynocomplex of the uppermost Pshekha Horizon (Samples 21 and 20-c, taken from 4.5 and 1 m below the Ostracoda Bed base respectively) is marked by a considerable increase in the Angiospermae pollen, in particular when approaching the Ostracoda Bed, thus reflecting the start of the warming period which was manifested even more distinctly during the accumulation of Solenovskii Horizon. In the Gymnospermae spectrum, comprising 60% of the entire complex, Taxodiaceae pollen predominates; whereas the pollen of *Angiospermae* shows predominance of the oak, sumach and *Tricolporopollenites* (>10%) and to a lesser extent of *Carpinus, Ilex, Comptonia, Engelhardtia* and *Carya*. The quantity of *Betulaceae* pollen decreases, whereas the pollen of the broad-leaved thermophilic species, namely, *Moraceae, Fothergilla, Liquidambar, Cyclocarya, Ilex, Nyssa, Elaeagnus*, and *Castanopsis*, is more diverse though scarce.

Phytoplankton of that level is scanty but extremely diverse. Both zonal index species of the lower Oligocene, namely, *Wetzeliella gochtii* and *Phthanoperidinium amoenum*, are present there, but there are no cysts of the *Cyclonephelium* and *Canningia* groups, which are abundant in the phytoplankton complex of the overlying Ostracoda Bed. The taxons which characterize the normal salinity environment of the open sea, dominate, namely, *Areosphaeridium pectiniforme, Spiniferites ramosus, Milliododinium tenuitabulatum, Lejeunecysta hyalina*, and also green algae like *Crassosphaera, Cymatiosphaera, Ptanospermella*, and *"Lanjaria."* Rare specimens of other taxons also occur there: *Distatodinium tenerum, Hystrichokolpoma salacia, Heterosphaeridium* cf. *conjunctum, Impletosphaeridium, Palaeocystodinium golzowense, Batiacasphaera, Homotriblium, Cannosphaeropsis, Dapsilidinium pseudocolligerum, Deflandrea leptodermata, Selenophemphix nephroides, Gelatia inflata, Lentinia wetzelii*, and *Apteodinium emslandensis*. Species of *Thalassiphora* and *Deflandrea* genera, and representatives of the *Wetzeliella* group, practically became extinct when approaching the Ostracoda Bed.

Phytoplankton of the Ostracoda Bed, both in the Kuban' section and at the Stavropol' arch, is rather rich in taxons (more than 40 species). The leading group of that stratigraphic level is the series of *Cyclonephelium-Glaphyrocysta-Adnatosphaeridium* (+*Canningia*). The latter plays the decisive role in that phytoplankton complex. The representatives of the genera of the *Wetzeliella* and *Deflandrea* groups are extremely scarce. Such genera as *Impagidinium, Deflandrea*, cf. *Hexagonifera*, and *Paucisphaeridium* are abundant. The presence of rare specimens of *Hystrichokolpoma salacia, Spinifer-*

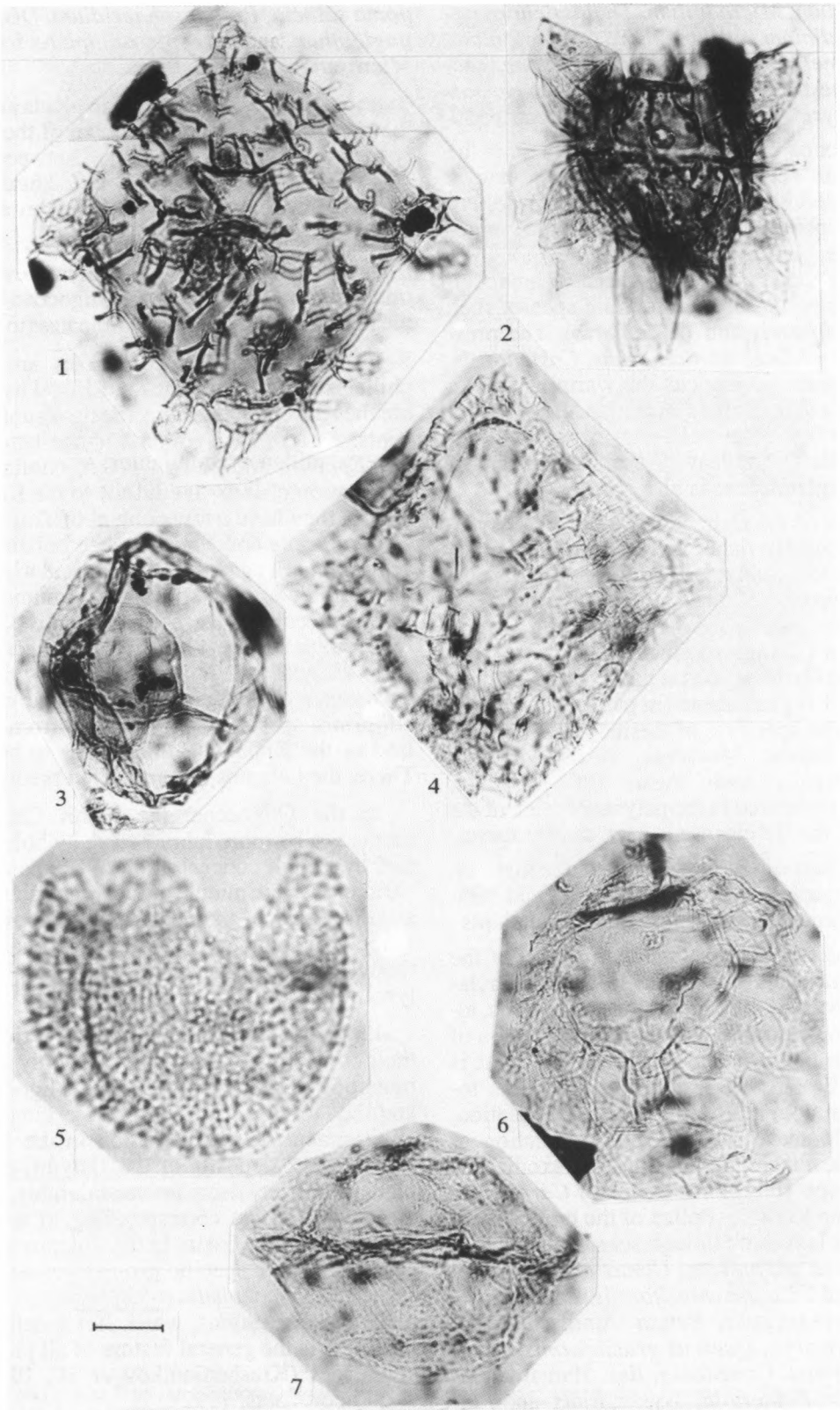


Plate I. *Phytoplankton of the Oligocene deposits of the Kuban' section, the scale for figure 5 is 10 μm, for all other figures – 20 μm. 1 – Wetzeliella gochtii Costa and Downie, Sample 23 - Kub/85, Pshekha Horizon; 2 – Hystrichokolpoma cinctum Klump, Sample 18-b -Kub/85, Belaya Glina Formation, 3 – Thalassiphora delicata Williams and Downie, Sample 23 - Kub/85, Pshekha Horizon, 4 – Charlesdownia clathrata (Eisenack) Lentin and Vozzhennikova, Sample 60 Ch/D, 4 - 7 – Ashcheairyk Formation, 5 – Baculisphaera baculata Drugg, Sample 22 Ashch., 6 – Adnatosphaeridium sp., Sample 22 Ashch., 7 – Heteraulacacysta porosa Bujak, Sample 21 Ashch.*

ites, Operculodinium, Microdinium, Thalassiphora pelagica, Gerlachidinium aechmophora, Leptodinium, Selenophemphix nephroides, Cleistosphaeridium, Tectatodinium, Rottnestia borussica, Eatonicysta, Homotriblium, Stoverocysta, and others has been recorded.

Palynocomplex of the Ostracoda Bed, which is the fourth in the Kuban' section, in contrast to the dinocyst one, differs considerably from that in the more northern region, the Stavropol' arch. In the former, the Gymnospermae dominate, particularly Taxodiaceae, though as in the northern section, the role of Fagaceae is appreciable (including the pollen of thermophilic species such as *Quercus graciliformis* and *Q. conferta*). The presence of the pollen of *Engelhardtia, Rhus, Castanopsis,* Palmae, and Moraceae evidences the warming during the accumulation of the Polba Formation. Besides Ulmaceae, there are few other pollen taxons of the temperate-thermophilic broad-leaved species; pollen of Betulaceae and Juglandaceae is also scarce.

Palynocomplex of the Ostracoda Bed of the Stavropol' arch is considerably richer and even more thermophilic, mainly due to the abundant occurrence of the pollen of Fagaceae (up to 50% of the total content) and, in particular, *Quercus graciliformis* (up to 40%). On the whole, the amount of Angiospermae pollen is twice as much as that of the Gymnospermae there. There is abundant appearance of Juglandaceae (in particular, *Carya*), Rhus, and Acer. The spectrum of thermophilic forms is diverse, though scarce: Moraceae, *Ilex, Magnolia, Palmae, Castanopsis,* and *Nyssa* (in particular, *N. granulata*). As compared to the palynocomplex of the Pshekha Horizon, the Betulaceae is even scarcer there.

Marine coefficient of the Ostracoda Bed is high: on the whole, the phytoplankton is as abundant and variable as the palynomorphic complex of the high plants.

The second half of the accumulation period of the Solenovskii Horizon, corresponding to the accumulation of the lower beds of the Morozkina Balka Beds, also remained warm. It is manifested in the proportion of the Gymnospermae and Angiospermae pollen, that is almost equal (52% and 48%), and a considerable increase of the thermophilic group pollen concentration, particularly of Moraceae (to 10%). Concentration of pine pollen is somewhat higher than that of Taxodiaceae (28% and 23% respectively); the pollen of *Carpinus* is rather abundant (up to 18%). Pollen of the typical early Oligocene species is present though scarce: *Carya glabraeformis, Juglans polyporata, Ulmus crassa, Ephedra cheganica,* and *Tilia tomentosiformis*. Pollen of *Picea, Cedrus, Glyptostrobus, Betula, Alnus, Corylus, Pterocarya, Cyclocarya, Quercus graciliformis, Q. sibirica, Fagus, Myrica, Comptonia, Ilex,* Hamamelidaceae, *Liquidambar, Fothergilla, Nyssa, Rhus,* and *Platanus* is rare. Spores are extremely scarce (Plate II).

Phytoplankton is scarce and scanty as compared to that of the Ostracoda Bed (up to 20 taxons). There, Spiniferites plays a considerable role, *Cyclonephelium* and *Canningia* are abundant, too, zonal index species *Wetzeliella gochtii* is present as well as *Hysterichokolpoma salacia, Prolixosphaeridium, Distatodinium, Impagidinium,* and *Areospharidium*. As for the acritarchs, *Micrhystidium* occurs there.

Palynocomplexes of the high plants and complexes of the organic-walled phytoplankton of the Kuban' section occupy the intermediate position between the complexes of the same age interval in Kazakhstan and Armenia (Krasheninnikov *et al.*, 1989; Akhmet'ev and Zaporozhets, 1992).

In those regions, sections analyzed evidence the maximum cooling at the Eocene/Oligocene boundary, when the Gymnospermae pollen dominates (to 90% and over).

The Kuban' palynocomplexes are more thermophilic as compared to those of Kazakhstan, and resemble the Armenian ones in variety of subtropical components. As a rule, in contrast to the latter, they contain less *Ephedra* and *Cedrus* pollen confined to the xerophytization phases. In contrast to the Kazakhstan complexes, they have a low content of *Tsuga* pollen as well as *Betulaceae* and Juglandaceae pollen, and high concentration of Fagaceae, and, particularly *Rhus*. In all regions studied, the warming maximum falls within the time of accumulation of the Solenovskii Horizon. Palynocomplexes show considerable decrease of amount and variety of Gymnospermae pollen with the abundant occurrence of Fagaceae, in particular of *Quercus graciliformis* species. Sample 20-b from the Ostracoda Bed of the Kuban' section seems to be an exception. There, the Gymnospermae pollen predominates.

In the Oligocene, the North Caucasian climate seemed to be more humid on the whole, which is verified by a high concentration of Taxodiaceae pollen, in particular, in the middle of the three Pshekha complexes. In all sections of the regions listed above, the Eocene–Oligocene boundary is marked by a change in the quantitative ratio of organic-walled phytoplankton and palynomorphs of the high plants in favor of the latter.

Phytoplankton assemblages are rich everywhere; their taxonomic diversity changes only slightly with the transition from the Eocene to the Oligocene in Armenia and the Northern Caucasus, preserving the normal sea environment of sedimentation in the basins. Only in Ashcheairyk deposits of the Ustyurt, does the phytoplankton variety decrease considerably, in particular, in the section range corresponding to the shoaling and freshening of the basin. In the Solenovskii Horizon, the appearance of a specific group of cysts, combined in a series *Cyclonephelium–Glaphyrocysta–Adnatospheridium* (+*Canningia*), which has a definite correlation meaning, is the general feature of all phytoplankton assemblages (Krasheninnikov *et al.*, 1989; Akhmet'ev and Zaporozhets, 1992).

In the regions studied, boundaries of the two early Oligocene dinocyst zones may be delineated reliably. These are *Phthanoperidinium amoenum* and *Wetzeliella gochtii* Zones. The *W. symmetrica* Zone is distinct only in the Ustyurt sections, where variations in taxonomic composition dinocyst facies may be identified

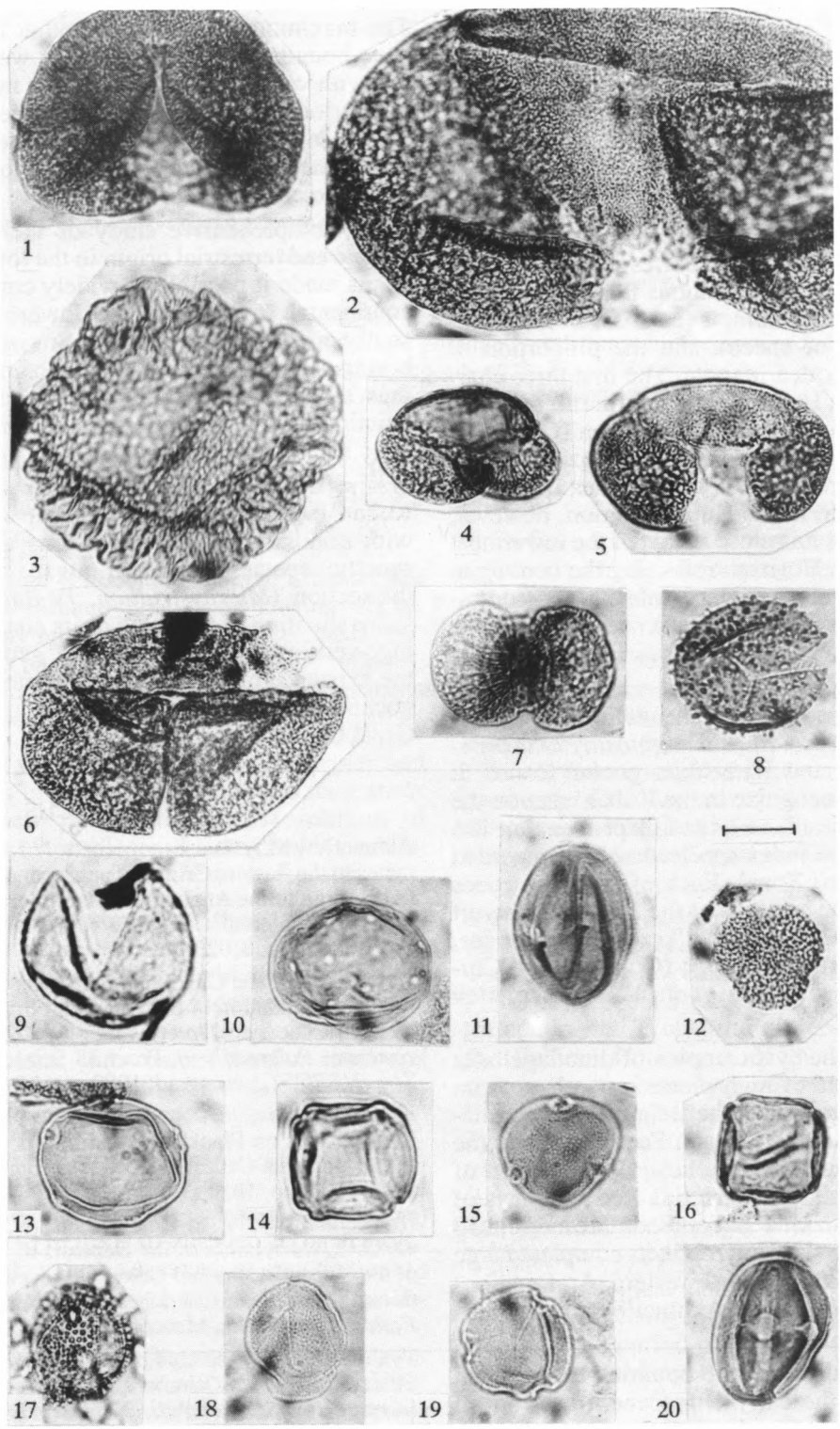


Plate II. *Pollen and spores from the Oligocene deposits of the Kuban' section: Ashcheairyk Formation. The scale for figures 9, 14 is 10 μ, for all other figures – 20 μm.*
1 – Cedrus sp., Sample 22 Ashch., 2 – Picea tobolica Pan., Sample 21 Ashch., 3 – Tsuga crispa Zakl., Sample 22 Ashch., 4 – Pinus sp., Sample 55 Ch/D, 5 – Pinus koraiensis Zakl., Sample 55 Ch/D, 6 – Picea alata Zakl., Sample 55 Ch/D, 7 – Podocarpus kazakhstanica Zakl., Sample 55 Ch/D, 8 – Osmunda sp., Sample 21 Ashch., 9 – Taxodiaceae, Sample 55 Ch/D, 10 – Juglans polyporata Vojcel, Sample 21 Ashch., 11 – Fagus sp., Sample 24 Ashch., 12 – Corylopsis crassa Lubm., Sample 24, 13 – Carpinus sp., Sample 21 Ashch., 14 – Alnus sp., Sample 55 Ch/D; 15 – Tilia grandireticulata Zhezhel., Sample 22 Ashch., 16 – Ulmus sp., Sample 24 Ashch., 17 – Ilex sp., Sample 27 Ashch., 18 – Hamamelis scotica Simpson, Sample, 19 – Nyssa granulata Pan., Sample 21 Ashch., 20 – Pokrovskaja originalis Boitzova, Sample 24 Ashch.

more precisely due to frequent changes of sea level and salinity.

Thus, reviewing the above information, one may arrive at the following conclusions:

1. The four subdivisions of the upper Eocene and lower Oligocene were studied in the reference section on the Kuban' River. They are Belaya Glina Formation, Pshekha Horizon, Ostracoda Bed, and the lower clayey part of the Morozkina Balka Beds. Four palynocomplexes were identified there, which correspond to the united temperate mesophyllic deciduous flora with certain subtropical ingredients. Complexes differ in the dominating components of spectra and the proportion of thermophilic plant pollen in them. The first three characterize the Pshekha Horizon, and the fourth, the most thermophilic one, the Solenovskii Horizon (Ostracoda Bed), and the lowermost beds of the Morozkina Balka Beds. There are scarce data on the complex of the Belaya Glina Formation in the Kuban' section, however, in its uppermost part, it is close to that of the lowermost parts of the Pshekha Horizon, reflecting the cooling at the Eocene/Oligocene boundary (absolute predominance of the Gymnospermae pollen).

2. Three zonal dinocyst complexes have been identified that characterize the transition from the upper Eocene to the Oligocene and the entire lower Oligocene: the *Wetzeliella clathrata angulosa*, *Phthanoperidinium amoenum*, and *Wetzeliella gochtii* Zones. It was not possible to recognize in the Kuban' section the *Wetzeliella symmetrica* Zone as an independent one (no first appearance of the index species has been recorded outside the *W. gochtii* Zone). Each of the three zones was recognized in the sections of the Northern Ustyurt basin, sections of the Southern Ukraine, Carpathians, and Central and Northern Europe (Gotsadze and Chikovani, 1989; Krasheninnikov *et al.*, 1989; Akhmet'ev and Zaporozhets, 1992).

3. According to the cyst complex of dinoflagellates and palynocomplexes of high plants, as well as from other groups of microfossils, the Belaya Glina Formation is correlated with the Chegan Formation, and the Pshekha Horizon with the Ashcheairyk Formation of the Northern Ustyurt. The Ostracoda Bed and clays of the lowermost Morozkina Balka Beds are correlated well with the Solenovskii Horizon in its complete range identified in hole 3006 (Northwestern Aral region) (Krasheninnikov *et al.*, 1989; Akhmet'ev and Zaporozhets 1992).

4. One may speak about the common tendency in climate changes of the early Oligocene, traced in all three regions studied (Northern Caucasus, Armenia, and Kazakhstan) (Akhmet'ev and Zaporozhets, 1992). The maximum cooling occurred at the Eocene/Oligocene boundary; then, the climate was getting warmer, but with variable humidity, as one may conclude from the periodic interchange of Pinaceae and Taxodiaceae taxons in the dominating group. The warmest climate of the interval studied corresponds to the accumulation of the Solenovskii Horizon.

5. Comprehensive study of the palynomorphs of marine and terrestrial origin in the lower Oligocene deposits made it possible to widely correlate marine and continental deposits of the lower Oligocene of the southern regions of the former Soviet Union and neighboring areas of the Russian craton and Turan platform, thus allowing the conclusions on climatic lateral restrictions of the major phytoplankton species.

6. The occurrence of organic-walled phytoplankton of the upper Eocene and lower Oligocene in the Kuban' section, which has great systematic variety with dominating open-sea taxons and practically no specific species of green algae and acritarchs in the section (*Micrhystridium*, *Pediastrum*), indicating sharp shoaling of the basin or its considerable freshening, verifies the conclusions of geologists who deny the existence of any vast area uplifted in the early Oligocene above sea level in place of the present-day Great Caucasus Ridge.

REFERENCES

Akhmet'ev, M.A. and Zaporozhets, N.I., The Oligocene Climates in the Eastern Crimea-Caucasian Area and Kazakhstan (According to the Analysis of Macroflora, Palynocomplexes and Phytoplankton Complexes), *Izv. AN SSSR, Ser. Geol.*, 1992, no. 3, pp. 5 - 23.

Gotsadze, V.I. and Chikovani, A.A., On the Peridinial Algae and Nannoplankton of the Khadum Formation in the Northwestern Caucasus, *Teoreticheskie i prikladnye Aspekty Sovremennoi Paleontologii*, Proc. 33 Session USSR Paleontol. Soc., VSEGEI, Leningrad, 1989, pp. 30 - 35.

Krasheninnikov, V.A. and Muzylev, N.G., Correlation of Zonal Scales on Plankton Foraminifers and Nannoplankton in the Northern Caucasus Sections, *Voprosy Mikropaleontologii*, 1975, no. 18, pp. 212 - 224.

Krasheninnikov, V.A., Zaporozhets, N.I., and Kurgalimova, G.G., Correlation of the Deposits of the Different Facies of the Upper Eocene and Lower Oligocene (Oceans and Continents), *Problemy Stratigrafii Verkhnego Proterozoya i Fanerozoya*, Nauka, Moscow, 1989, pp. 199 - 220.

Vyalov, O.S., On the Combined Subdivision of the Maikop Strata, *Maikopskie Otlozheniya i Ikh Vozrastnye Analogi na Ukraine i v Srednei Azii*, Naukova Dumka, Kiev, 1964, pp. 23 - 35.

Stratigraphy and Geological Correlation, Vol. 1, No. 1, 1993.Translated from Stratigrafiya i Geologicheskaya Korrelyatsiya, Vol. 1, No. 1, pp. 137 - 143.
Original Russian Edition Copyright © 1993 by Stratigrafiya. Geologicheskaya Korrelyatsiya, Fedorov.
English Translation Copyright © 1993 by Interperiodica.

The Pontian-Caspian Early Pleistocene

P. V. Fedorov

Geological Institute, Russian Academy of Sciences, Pyzhevskii Per. 7, Moscow, 109017 Russia

Manuscript received April 16, 1992

Abstract – The biostratigraphic subdivision of the Bakuan succession (the Caspian) and the Chaudan succession (the Black Sea) is studied for comparison and possible correlation with the corresponding strata in the Mediterranean Sea. The sequences of the Chaudan deposits on the Kerch' Peninsula and in the Sea of Azov area are compared to the Guriyan section, with due allowance for paleomagnetic data.

The geological history of the Caspian and Black seas is, as yet, inadequately studied, particularly the early Pleistocene in the Black Sea region (Table 1)[1].

Since Kimmerian (Pliocene) time, the Caspian Sea developed as an isolated lake basin casting off its waters periodically into the Azov-Black Sea basin, and the Black Sea was an inland sea whose history reflected the main stages in the evolution of the Mediterranean and the Caspian seas.

The history of the unique basins of the Caspian and Black seas is rich in geological events, and their study may open up new possibilities in the solution of regional and general problems of stratigraphic correlation and paleogeography of southeastern Europe.

The changes in the hydrology of the Black Sea basin between the Pliocene and the Eopleistocene have caused a renewal of its link with the Mediterranean. For example, in the Pliocene - Eopleistocene (Kimmerian, Kuyal'nikian, and Guriyan), the water flowed into the Mediterranean basin, but since Chaudan time and throughout the Pleistocene, the Mediterranean waters repeatedly flooded into the Black Sea basin.

There were several major events in the Early Pleistocene in the Black Sea-Caspian region. In the end of the Eopleistocene, the post-Apsheronian (Tyurkyan) regression occurred in the Caspian Sea. Tyurkyan alluvial and alluvial deltaic deposits were discovered between the Apsheronian and Bakuan sediments by drilling in the Kura Lowland, on sea bottom, in Baku Archipelago, in northern Dagestan, in the lower reaches of the Volga, and in western Turkmenistan. These deposits are now exposed in anticlinal structures on the northern side of the Kura basin and in western Turkmenistan; they become gradually thinner toward the Caspian basin and then wedge out. Everywhere the Tyurkyanian deposits overlie the eroded Apsheronian and older rocks and grade into Bakuan sediments. The presence of ostracodes of the Bakuan type identifies them as the lowermost beds of the Bakuan Horizon.

[1] The Eopleistocene is not discussed here.

In the Black Sea area, the evidence of the post-Pliocene regression is less obvious.

In Guriya (western Georgia), between the Chaudan sediments and the underlying Guriyan sequence, there are the transitional (Natanebian) beds, containing mostly the Guriyan molluscan fauna and rare ancestral forms of the Chaudan fauna.

Recent research on the shelf of Bulgaria and in the southwestern Ukraine has shown the Chaudan succession overlying the Guriyan (Dimitrov and Govberg, 1979; Krystev *et al.*,1990; Kuprin *et al.*, 1984). In some places, the Chaudan marine deposits overlie continental rocks, thus indicating an interruption in sedimentation between deposition of the Guriyan and Chaudan.

Undoubtedly, at the Pliocene/Eopleistocene boundary, a regression occurred in the Black Sea region as well, but it left less evidence than in the Caspian basin. The stratigraphic constitution of the Caspian Lower Pleistocene beds is rather complicated. If we assume the continental Tyurkyan Formation as the basement of the Bakuan Horizon, then the Matuyama/ Brunhes paleomagnetic boundary lies at the base of the lower Bakuan marine beds (or in the upper beds of the Tyurkyan Formation), and the Jaramillo event can be traced in the upper Apsheronian. Consequently, all Bakuan deposits fall within the Brunhes epoch, and the Tyurkyan Formation correlates with the upper part of the Matuyama epoch.

The geological data and tectonic inconformities, and a sudden change in the evolution of molluscs, indicate that the period of Tyurkyanian regression was a prolonged and important event in the history of the Caspian Sea.

In the end of the Eopleistocene, universal climatic changes apparently caused the aridization of climate and a considerable regression in the isolated Caspian. The widespread continental red rocks in the northern part of the Caspian basin and in the Black Sea area, in particular the Scythian clays overlying the Apsheron Formation, bear evidence to the desiccation and warming of the climate. The characteristic features of the Tyurkyan deposits imply an abundant supply of river

Table 1. Geological correlation of the Caspian, Black and Mediterranean seas in the Pleistocene

ISC Scale		The Mediterranean Sea (the author's interpretation of data from the literature)		The Black Sea (after the author's data and published materials)		The Caspian Sea (after the author's data)	
Holocene		Flandrian (maximal phase)		Chernomorian		Neocaspian	
Pleistocene	Upper	Flandrian (first phase)		Neoeuxinian (transgression from 90 to 30 m below sea level)		Khvalynian	Upper Khvalynian
		Grimaldian regression (120 - 130 m below sea level)		Post-Karangatian (regression to 90 m below sea level)			Lower Khvalynian
						Atelian regression	
		Tyrrhenian	Chebba	Karangatian	Upper (main) Karangatian	Upper Khazarian	
			Rejiche			Regression	
			Douira		Lower Karangatian	Lower Khazarian	Lower Khazarian - III (Kosogian)
	Middle	Regression		Regression			Regression
		Transgression (Mesotyrrhenian)		Euxinian-Uzunlarian	Uzunlarian		Lower Khazarian - II (Singilian)
		Regression			Upper Old Euxinian		
		Paleotyrrhenian			Regressive phase		Regression
					Paleouzunlarian		
	Lower	Roman regression (Mindel)			Lower Old Euxinian		Lower Khazarian - I (Paleosingilian)
				Venedian (Roman) regression		Venedian (Ushtalian) regression	
		Millazzian (Warm Cromerian)		Chaudan	Epi-Chaudan		
					Upper Chaudan (Cape Chauda, Caucasian coast)	Bakuan	Urundzhik
							Upper Bakuan
		Regression (Günz)			Lower Chaudan (Cape Chauda, northern coast of Taman' Peninsula)		Lower Bakuan
					Bulgarian Chaudan (regression, shelf)		Tyurkyan regression
					Guriyan Chaudan (?)	Apsheronian	
Eopleistocene		Calabrian		Guriyan			

and melted glacial water from the mountains of Central Asia and the Caucasus. For example, the Tyurkyan rocks recovered from holes drilled near Kizlyar, Chernyi Rynok, and Krainovka in northern Dagestan contain large cobbles and boulders (up to 10 or 15 cm in diameter judging by fragments) of crystalline and metamorphic rocks from the Main Caucasus (Fedorov, 1978).

The old coarse-sand and granule gravel alluvium of the paleo-Amudar'ya, which corresponds to the Tyurkyan Formation, is exposed in the dome structures of western Turkmenistan. As evidenced by the palynological data, the lower Bakuan deposits accumulated at a later period in a more humid climate and are now represented in all parts of the basin by the typical laminated brownish-grey clay and silt, with thin interbeds of ocherous sand.

The general desalting of the basin, due to greater inflow of water from the mainland and the growth of water mass during the early Bakuan transgression, is confirmed by the small and thin shells of molluscs *Didacna parvula* Nal. and *D. catillus* Eichw., and by the character of an ostracode assemblage.

A considerable thickness of the lower Bakuan deposits (300 - 400 m and sometimes more) and their rather monotonous lithology within the entire basin imply a long and extensive early Bakuan transgression, which filled all topography lows and from the Manych valley spilled out into the basins of the Sea of Azov and the Black Sea. After an interval (possibly a regressive phase), the upper Bakuan deposits started to accumulate during a later, rather short transgression of the sea, which perhaps exceeded the level and area of the earlier phase.

The upper Bakuan deposits, observed in natural outcrops and recovered from drilled holes, are composed of coarse coquinas, conglomerates, sands, and seldom silts, from several meters to several tens of meters thick. They normally contain massive shells of *Didacna rudis* Nal., *D. carditoides* Andrus., and less often *D. eulachia* (Bog.) Fed. The character of sediments, the amount of carbonate cement in them, and the thickness of shells may indicate rather high temperatures of water rich in calcium salts. The final transgression is responsible for the deposition of the Urundzhik beds, represented by sands and coquinas, and less often by silts several meters thick with large shells of *Didacna eulachia* (Bog.) Fed., *D. pravoslavlevi* Fed., *D. kolesnikovi* Fed., and *D.* ex gr. *nalivkini* Wass.

D. eulachia first appears in the upper beds of the lower Bakuan deposits (the Mishovdag section, the middle part of the Neftyanaya Balka section). It is often found in the upper Bakuan sediments, and with the forms listed above it dominates in Urundzhik beds succession registered in western Turkmenistan (Urundzhik), on Cape Cheleken, and at Mishovdag (Nefandtyanaya Balka) in the Kura Lowland (Fedorov, 1978).

On the slopes of the southeastern Caucasus, these strata compose an independent abrasion-constructional terrace 180 - 200 m high, below the upper Bakuan terrace (220 - 230 m). Urundzhik shells were observed by G.I. Popov (1983) in the northwestern Caspian, and in the Sea of Azov area; they are also present in the upper Chaudan beds on Cape Chauda (Fedorov, 1963; Fedorov, 1979).

In the Karadzha and Duzdag brachyanticlinal folds in the northern Kura Lowland, the upper Bakuan and Urundzhik sediments are overlain by a layer of rosy-colored volcanic ash from 1 to 2 m thick and about 0.5 - 0.6 Ma old (age determined by fission-track method). This period apparently correlates in the Pontian-Caspian regions with the final stage of the early Pleistocene, which corresponds to the end of the global warming of the climate just before the Mindel (Oka) Glaciation.

This glaciation caused a great regression of the World Ocean, of the Mediterranean, and of the Black Sea (the Roman or Venedian regression). The evidence of the Venedian regression in the Black Sea basin is documented by drilling in the lower reaches of the Don, Dnieper, and other rivers (Goretskii, 1966). The Venedian deposits and their analogs (the Mindel, in general) are traditionally referred to as the lower Pleistocene. The properties of these sediments and their paleontological characteristics, however, are indicative of the principally new physicogeographical conditions that have set in after the termination of Bakuan-Chaudan cycle of sedimentation. It is the turning point in the evolution of the brackish-water molluscs, and several ancestral forms of the Khazarian time appear at the very end of the Bakuan.

On the strength of these data, the Venedian Formation (and the Roman regression) should be transferred to the base of the middle Pleistocene, dating it at about 0.4 - 0.5 Ma.

The lower Pleistocene of the Caspian (or the Bakuan Horizon) includes the continental Tyurkyan Formation, the lower Bakuan, the upper Bakuan and the Urundzhik beds; in the paleomagnetic scale it covers the period from the uppermost Matuyama epoch to the lower Brunhes epoch and most likely corresponds to the time interval from 0.8 (or somewhat more) to 0.5 Ma ago.

It should be noted that the long development of the Bakuan transgression in humid climate (the Early Bakuan) has produced a vast basin, which later existed in a warm but probably not arid climate (late Bakuan). At the concluding evolutionary stage of the basin (Urundzhik time), the climate became more humid.

The Black Sea basin had a far more complicated history, but its study is hampered by an incomplete geological record: there are no complete natural sections showing the succession of sedimentation, and the drilling data on the mainland and on the shelf do not always contain the typical fauna.

Deposits with the typical Chaudan molluscan fauna have been found only at a few places on the Caucasian

coast (Guriya, Tuapse, Cape Idukopas) and in the southern part of the Kerch' Peninsula (Cape Chauda); recently, they were also found on the Black Sea shelf (Dimitrov and Govberg, 1979; Krystev *et al.*, 1990a; Kuprin *et al.*, 1984), Therefore, we should be very cautious when studying the Early Pleistocene records of the basin.

Since N.I. Andrusov's research, the section on Cape Chauda was assumed as a stratotype (Andrusov, 1965). The Chaudan deposits overlie eroded Maikopian rocks, which are more than 10 m above sea level. They are studied in several papers (Fedorov, 1963, 1978).

The lowermost Chaudan beds are composed of sands, gravels, and silts, 3 - 4 m thick, and contain *Didacna baericrassa* Pavl., *D. parvula* Nal., *D.* aff. *pallasi* Prav., *D.* aff. *olla* Liv., *Monodacna subcolorata* Andrus., *Dreissena caspia* Eichw., *Dr. polymorpha* Pall., *Dr. tschaudae* Andrus., *Viviparus pseudoachatinoides* Pavl., *Micromelania caspia* Eichw., and *Pisidium* sp. The first two forms and *Dreissena* and *Viviparus* compose the bulk of the Bakuan (or Chaudan-Bakuan) fauna in the deposits of the terrace on the northern coast of the Taman' Peninsula, including Cape Akhillion (Cape Litvinov).

Continental loams, 2 - 3 m thick, above these layers show traces of soil at the top. They are unconformably overlain by stratified limestones (coquina, loose sandstones) 0.5 - 4 m thick with a reach assemblage of Chaudan fossils. They are most numerous in the beds of loose sandstones including the uppermost layer, which is weathered out into sands. They contain the following fossils: *Tchaudia tschaudae* (Andrus.), *Didacna pseudocrassa* Pavl., *D. rudis* Nal., *D.* cf. *eulachia* (Bog.) Fed., *D. carditoides* Andrus., *D. plesiochora* Davit., *D. pleistoplura* Davit., and *D. olla* Livent. Less numerous are *Didacna* aff. *subpyramidata* Prav., *D. pallasi* Prav., *Monodacna cazecae* Andrus., *Dreissena polymorpha* Pall., *Dr. tschaudae* Andrus., and *Theodoxus pallasi* Lindh. Many shells are covered with yellow-brown carbonate incrustations of calcium carbonate as a result of its high concentration at rather high temperatures (Fedorov, 1979). It is important to note that the incrustations cover both the Pliocene relics *Tschaudia*, *Didacna pleistopleura*, etc. and *D. pseudocrassa*, *D. rudis*, etc. All these shells occur *in situ*.

A comparison of these forms with the fauna from the Chaudan beds in Guriya, which lie above the Natanebian Horizon, reveals their doubtless similarity. Moreover, on Cape Chauda, an important role belongs to the Bakuan immigrants: *Didacna parvula* Nal., *D. carditoides* Andrus., and *D. rudis* Nal., whereas in Guriya the rocks sometimes contain local forms: *Tschaudia guriyana* Livent., *Didacna crassa supsae* Newessk., etc. It should be noted that on Cape Chauda *Didacna carditoides* Andrus., *D. rudis* Nal., and *D. eulachia* (Bog.) Fed. are similar to the analogous forms from the upper Bakuan and Urundzhik deposits of western Turkmenistan and Azerbaidzhan. In the Chaudan deposits of Guriya, I often found, as some researchers have, the typical *Didacna baericrassa* Pavl., which is characteristic of both Cape Chauda and the Caucasian coast.

The data given above still show no evidence that these deposits from Cape Chauda and from Guriya were formed synchronously, because according to paleomagnetic data the former correspond to the Brunhes epoch and the latter to the Matuyama epoch. It seems obvious that the molluscs of the Guriyan section inhabited the basin since the very beginning of Chaudan time, being genetically connected with the ancestral forms from the underlying Natanebian deposits. The presence of analogous shells in the upper Chaudan beds on Cape Chauda is evidence that the Chaudan transgression advanced there at a later date, and it seems plausible to superpose the Cape Chauda section over the Guriya section.

Consequently, accumulation of the lower Chaudan layers of the Black Sea started even earlier than 0.7 Ma ago and terminated, apparently, about 0.5 Ma ago, as evidenced by a comparison between the upper Chaudan deposits on Cape Chauda and the upper Bakuan beds in the Caspian sequence, including the Urundzhik (Fedorov, 1979).

Let us now consider the correlation between the deposits on Cape Chauda and on the northern coast of the Taman' Peninsula, where a terrace about 35 - 40 m high can be traced from Cape Akhillion to Cape Pekla; it is composed of light-yellow sands overlying the deformed Sarmatian rocks. These beds are covered by loams with several interbeds of buried soils (Fedorov, 1978). The sands contain the Bakuan type molluscs *Didacna parvula* Nal., *D. baericrassa* Pavl., *D. catillus* Eichw.; less numerous are *D. catillus-rudis* Nal., *D. pseudocrassa* Pavl., *Dreissena caspia* Eichw., *Viviparus pseudoachatinoides* Pavl., *Margaritifera moldavica* Tschep., and others. The dominant forms are *Didacna parvula* Nal. and small *D. catillus* Eichw., which are typical of the lower Bakuan deposits in the northwestern Caspian coasts and in the lower reaches of the Volga.

A similar fauna is present in the Bakuan sands occurring at the base of the coastal section at Malyi Kut Hamlet in the western part of the Taman' Peninsula (Fedorov, 1963). Moreover, in the upper part of the Manych Valley, at a depth of 40 m below sea level, the Bakuan deposits were found to contain *Didacna rudis* Nal. And finally, on the northern coast of the Sea of Azov near the mouth of the Mius Liman, the deposits of a terrace, known in literature as the Platov's terrace, contain rather thin-valved, large, flat *Didacna* cf. *carditoides* Andrus., *D.* cf. *eulachia* (Bog.) Fed., and *D.* cf. *nalivkini* Wass. The top of these beds at the terrace's brow is 9 - 10 m above sea level; they are covered by intricately built continental loams.

G.I. Popov (1983) and N.A. Lebedeva believe that the Platov's deposits belong to the lower Bakuan. Previously, I dated them at an earlier period corresponding to the lower Old Euxinian or the Urundzhik (Fedorov,

1978). Repeated studies of the mentioned fossils and their comparison with the possible analogs from the lower Pleistocene of the Caspian and the Black Sea show, however, that most likely they correlate with the Urundzhik beds or with the uppermost Bakuan Horizon, that is, with the end of the early Pleistocene.

What then is the stratigraphic correlation of the studied deposits on the coasts of the Sea of Azov? No doubt, the oldest beds, the lower Bakuan, are present on the northern coast of the Taman' Peninsula. Here they are uplifted, and their position does not correspond to the former sea level. The Platov's layers are perhaps somewhat uplifted too, but at least not lowered. The position of the Bakuan deposits in the Manych Valley may indicate the low base level of erosion. The Bakuan deposits at Malyi Kut Hamlet have probably subsided.

The paleontological data and the extent of tectonic deformations provide grounds to state that the Sea of Azov basin during the first half of the early Pleistocene was a running brackish-water body with a water supply from the early Bakuan Caspian. Along a channel, which did not exactly coincide with the modern Kerch' Strait, the water drained into the Black Sea where the Chaudan basin had a lower level. Traces of this drainage are, probably, the lower Chaudan gravels, sands, and clays with *Didacna parvula* Nal. and other fossils in the Cape Chauda section. These deposits definitely indicate the low level of the Black Sea. It can be inferred that during this time the Chaudan molluscs, whose remains can still be found in the Guriyan section, inhabited the Black Sea basin. Migration of this fauna into the Sea of Azov was prevented by its desalting and by the difference in sea levels.

The drilling data from the western shelf of the Black Sea confirmed the wide distribution of the Chaudan deposits in that area (Krystev *et al.*, 1990a, 1990b; Kuprin *et al.*, 1984; 1990b).

The initial information about the Chaudan deposits on the shelf of Bulgaria was obtained by P.S. Dimitrov and L.I. Govberg (1979) from the 80 to 100 m isobath interval. It was later supplemented by new findings which show that the Chaudan clays and clayey sands with typical molluscan fauna cover considerable areas of the shelf shallows from 100 to 80 - 40 m deep. The coastal facies of these deposits could also have extended on the mainland but later were washed out. Therefore, the Chaudan age of the 100-m abrasion terrace on the coast of Bulgaria remains an open issue (Fedorov, 1963). On the shelf of Bulgaria, the Chaudan deposits can be divided into two parts (Kuprin *et al.*, 1984): the lower beds are composed mostly of clayey sands, sometimes with pebbles, 5 - 7 m thick, and the upper layers, 3 - 4 m thick, are composed of clays with a rich assemblage of large and often massive shells of *Tschaudia tschaudae* (Andrus.), *Didacna pseudocrassa* Pavl., *Dreissena tschaudae* Andrus.; the lower beds contain also *Didacna pleistopleura* Davit. The specific composition and the look of the shells match well with the forms in the Guriyan section and in the upper Chaudan beds on Cape Chauda, and this resemblance justifies their correlation.

The modern position of the Chaudan deposits on the outer shelf implies a comparatively low sea level, but the lithology of the upper Chaudan here indicates depths of more than 50 m marked by finely elutriated clays. It is obvious that the lower Chaudan beds were deposited when the sea level was even lower, i.e., at the very beginning of the early Chaudan transgression.

A more detailed study of the section on Cape Chauda shows that if the Chaudan deposits there are younger than the beds of the same name in Guriya, then the Chaudan fauna of molluscs was already inhabiting the main basin earlier than the paleomagnetic Matuyama/Brunhes boundary.

The advancing transgression started to fill with deposits the erosion hollows of the paleoKerch' Strait, and the lowered portion of the base of the lower Chaudan deposits in the section on Cape Chauda is one of the relics of this paleostrait. Because of the slightly salted waters flowing from the Azov depression, the ecology of the strait was unfavorable for the Chaudan fauna of the Black Sea basin. At that time, a rather deep bay was situated in the place of modern Guriya upland, and clayey sediments deposited there contained abundant remains of the rich Chaudan fauna. The developing transgression advanced the coastline into the Azov area, transported this fauna to northeast, and flooded the south of the Kerch' Peninsula.

The continental deposits on Cape Chauda occurring between the sands and silts containing the lower Bakuan fauna and the upper coastal marine coquina with numerous remains of the Chaudan fauna indicate a certain break in sedimentation. It implies that before the invasion of the Chaudan sea the waterflow in the strait dried up, and here, as on adjacent areas, the cover deposits began to form. This could be caused by the uplift of a part of the Kerch' Peninsula and shifting of the stream to another watercourse (probably eastward). As noted above, the high position of the Paleogene clays is an indication of this uplift.

In Guriya, in contrast to Cape Chauda, sedimentation of the Chaudan (like Guriyan) clays was apparently continuous, and only at the end of Chaudan time the uplifts brought these deposits above sea level.

On Tsvermagal mountain in the Guriya upland, at an altitude of about 150 m, a bed of ocherous sands overlying the Chaudan deposits contains sparse remains of marine molluscs represented by *Cardium* cf. *edule* L. (*Cerastoderma glaucum*), *Scrobicularia* cf. *plana* (Costa), and *Paphia* sp. I referred these sands to the uppermost Chaudan (Fedorov, 1982) as the epi-Chaudan beds[2]. More recently, V.V. Yanko (1989) after detailed studies has established that the uppermost Chaudan deposits in the high terrace of the Caucasian coast (Cape Idukopas) and in the Cape Chauda section

[2] A.L. Chepalyga calls them Karadenizian.

contain marine benthonic foraminifers which inhabited waters with higher salinity.

The marine layers with the Mediterranean fauna overlying the deposits with the Chaudan fauna were recorded by Andrusov as early as 1904 in Dardanelles near Gallipoli (Andrusov, 1965), and at a later date more completely by B. Keraudren (1979) at the same place and near Murefte-Chora. Moreover, Keraudren points out the gradual replacement of the brackish-water fauna by the marine fauna. This evidence suggests salinization of the Black Sea during the maximum of the Chaudan transgression when the Mediterranean waters flowed through the Sea of Marmora.

It was the first penetration of marine fauna into the Black Sea area since Pontian time, and it marked the appearance of new characteristic features in the Black Sea evolution for the whole Pleistocene. At an early stage, these new features involved a consecutive development of the low-salinity transgressions associated with an abundant inflow from rivers and from the Caspian in late glacial and postglacial time; at a later period, the transgressions were caused by backwater of the Mediterranean during the maxima of global interglacial transgressions. The backwater was already developing, no doubt, during the low-salinity (Caspian) phase of transgression.

Similar events can be traced for the first and second Euxinian-Uzunlarian transgressions and even more distinctly in the Neoeuxinian (low-salinity) and Neochernomorian (Mediterranean) transgressions. It is now obvious that the course of events in the early Pleistocene was the same, and the level of the Chaudan basin reached its maximum at the end of the transgression. However, it is unlikely that the Chaudan fauna could have penetrated into the Azov basin, because its northern limit is located at Cape Chauda and, perhaps, near Cape Tuzla in the southwestern part of the Taman' Peninsula where coastal conglomerates contain rounded fragments of Kimmerian shells and the Chaudan *Tschaudia tschaudae* Andrus. (Fedorov, 1963). Only some late Chaudan (and late Bakuan) forms could find ecological niches. Therefore, the deposits of the Platov's terrace on the northern coast of the Sea of Azov may correspond just to that moment of transgression.

The presence of the Chaudan deposits on the coasts of the Sea of Marmora and the Dardanelles suggests that the Chaudan low-salinity waters may have penetrated into the Aegean area also immediately before the maximum of the Mediterranean interglacial transgression.

The global interglacial transgressions, considered against the background of the Black Sea history, show that they occurred possibly in the second half of the early Pleistocene (late Chaudan and EpiChaudan), twice in the middle Pleistocene (Paleouzunlarian and Uzunlarian), in the beginning of the late Pleistocene (Karangatian), and in the middle of the Holocene (Chernomorian).

The two latter transgressions obviously correlate with the Tyrrhenian and the Flandrian (the maximal phase), whereas the earlier transgressions require additional substantiation of their correlation with the Paleotyrrhenian and warm Cromerian (Millazzian).

The paleomagnetic data testify that the Chaudan deposits in Guriya correspond most likely to the Tyurkyan Formation and the upper Apsheronian, because the Jaramillo Event is recognized in their lower beds. This observation indicates that the Chaudan in Guriya is older (at least its lower beds) than the lower beds of the Baku Horizon.

It could be then assumed that initially the appearance of the PontianCaspian fauna of *Didacna* Eichwald occurred in the southeastern part of the Black Sea basin. Geological evidence from the Manych Valley, however, contradicts the possibility of migration of this fauna into the Caspian (Goretskii, 1966; Popov, 1983). An opposite direction of migration is definitely indicated there with the runoff of the Caspian waters into the Black Sea. The composition of fauna in the Chaudan deposits on Cape Chauda, in the high terrace of the Caucasian coast, and in the upper Chaudan in Guriya, where the Bakuan forms are of essential importance is no less convincing. It is interesting to note that in the late Pliocene the waters from the Caspian were likewise discharged into the Black Sea basin (the Akchagylian-Kuyal'nikian, the Apsheronian-Guriyan).

The supposition that at the very end of the Pliocene, after the Apsheronian waters were spilled into the Guriyan basin, there was a short-term migration of Chaudan cardioids into the Caspian seems unjustified. It is more appropriate to suggest parallel development of these faunas in the two basins.

The problems discussed here need more comprehensive research primarily on the shelf and continental slope of the Black Sea and the Caspian Sea.

REFERENCES

Andrusov, N.I., *Izbrannye Trudy* (Selected Works), Nauka, Moscow, 1965, vol. IV (1904 - 1905), pp. 143 - 162.

Dimitrov, P.S. and Govberg, L.I., New Data on Pleistocene Terraces and Paleogeography of the Bulgarian Shelf of the Black Sea, *Geomorfologiya*, 1979, no. 2, pp. 81 - 89.

Fedorov, P.V., Stratigraphy of Quaternary Deposits on the Crimean-Caucasian Coasts and Certain Problems of Geological History of the Black Sea, *Trudy GIN AN SSSR*, 1963, issue 88.

Fedorov, P.V., The Pontian-Caspian Pleistocene, *Trudy GIN AN SSSR*, 1978, no. 310.

Fedorov, P.V., Development of Molluscs of Didacna Eichwald Genus in the Pontian-Caspian Lower Pleistocene and Their Stratigraphical Implications, *Izv. AN SSSR, Ser. Geol.*, 1979, no. 3, pp. 79 - 87.

Fedorov, P.V., Some Debatable Problems of the Black Sea Pleistocene, *Byul. MOIP, Otd. Geol.*, 1982, vol. 57, no. 1, pp. 108 - 117.

Fedorov, P.V., Changes in the Level of the Black Sea in the Pleistocene, *Byul. MOIP, Otd. Geol.*, 1988, vol. 63, no. 4, pp. 55 - 61.

Goretskii, G.I., *Formirovanie Doliny R. Volgi v Rannem i Srednem Antropogene* (Formation of the Volga Valley in the Early and Middle Anthropogene), Nauka, Moscow, 1966.

Keraudren, B., Le Plio-Pleistocene Marin et Oligohalin en Grèce, *Rev. Geol. Dynam. et Geogr. Phys.*, 1979, vol. 21, pp. 17 - 28.

Kuprin, P.N., Samsonov, A.I., Babak, E.V., Varushchenko, A.N., Monakhov, I.B., and Fedorov, P.V., Structure and Biostratigraphical Subdivision of the Quaternary Deposits on the Shelf of Bulgaria, *Byul. MOIP, Otd. Geol.*, 1984, no. 3, vol. 59, pp. 31 - 40.

Krystev, T.I., Limonov, A.F., Sorokin, V.M., *et al.*, *Geologicheskaya Evolyutsiya Zapadnoi Chasti Chernomorskoi Kotloviny v Neogen-Chetvertichnoe Vremya* (Problems of the Chaudan deposits on the Bulgarian Black Sea Shelf), Bulgarian Academy of Sciences, Sofia, 1990a, pp. 349 - 361.

Krystev, T.I., Grigor'ev, A.V., and Fedorov, P.V., *Geologicheskaga Evolyutsiya Zapadnoi Chasti Chernomorskoi Kotloviny v Neogen-Chetvertichnoe Vremya* (The Pleistocene Biostratigraphy of the Bulgarian Black Sea Shelf and the Transgression - Regression Cycles of the Black Sea), Bulgarian Academy of Sciences, Sofia, 1990b, pp. 340 - 348.

Popov, G.I., *Pleistotsen Chernomorsko-Kaspiiskikh Prolivov v Pleistotsene* (The Black Sea-Caspian Straits in the Pleistocene), Nauka, Moscow, 1983.

Yanko, V.V., *Chetvertichhye Foraminifery Ponto-Kaspiya (Klassifikatsiya, Ecologiya, Stratigrafiya, Istoriya Razvitiya)* (The Quaternary Foraminifers of the Pontian- Caspian (Classification, Ecology, Stratigraphy, Evolution)), Doctorate (Geol.-Min.) Dissertation, Moscow State University, Moscow, 1989,.

Subscribe to the journals from the Russian Academy of Sciences and INTERPERIODICA

Copy this form to mail or fax your order today

TITLE	ISSN	ISSUES	RATE IN US DOLLARS US/CAN	RATE IN US DOLLARS INTERNATIONAL	YOUR ORDER
❑ **Geology of Ore Deposits**, Vol. 35	0016-7770	6	$ 540	$ 576	$_____
❑ **Glass Physics & Chemistry**, Vol. 19	Pending	6	$ 715	$ 751	$_____
❑ **Herald of the Russian Academy of Sciences**, Vol. 63	1019-3316	12	$ 315	$ 387	$_____
❑ **High Temperature**, Vol. 31	0018-151X	6	$1075	$1260	$_____
❑ **Inorganic Materials**, Vol. 29	0020-1685	12	$1275	$1490	$_____
❑ **Kinetics and Catalysis**, Vol. 34	0023-1584	6	$1175	$1375	$_____
❑ **Laser Physics**, Vol. 3	1054-660X	6	$ 400	$ 436	$_____
❑ **Pattern Recognition & Image Analysis**, Vol. 3	1054-6618	4	$ 240	$ 264	$_____
❑ **Petrology**, Vol. 1	0869-5911	6	$ 540	$ 576	$_____
❑ **Polymer Science**, Vol. 35	0965-545X	12	$1680	$1752	$_____
❑ **Russian Journal of Coordination Chemistry**, Vol. 19	Pending	12	$ 795	$ 930	$_____
❑ **Russian Journal of Electrochemistry**, Vol. 29	Pending	12	$1195	$1400	$_____
❑ **Russian Journal of Inorganic Chemistry**, Vol. 38	0036-0236	12	$1590	$1670	$_____
❑ **Russian Journal of Physical Chemistry**, Vol. 67	0036-0244	12	$1590	$1670	$_____
❑ **Russian Journal of Plant Physiology**, Vol. 40	Pending	6	$1075	$1260	$_____
❑ **Stratigraphy and Geological Correlation**, Vol. 1	0869-5938	6	$ 540	$ 576	$_____
❑ **Studies on Russian Economic Development**, Vol. 4	Pending	6	$ 230	$ 266	$_____
❑ **The Physics of Metals and Metallography**, Vol. 75/76	0031-918X	12	$1386	$1458	$_____
❑ **Theoretical Foundations of Chemical Engineering**, Vol. 27	0040-5795	6	$ 995	$1165	$_____
❑ **Thermal Engineering**, Vol. 40	0040-6015	12	$ 875	$ 904	$_____
❑ **Water Resources**, Vol. 19	0097-8078	6	$ 765	$ 895	$_____
				TOTAL	_____

❶ Name _____

Institution _____

Address _____

_____

City _____ State _____

Zip/Postal Code _____ Country _____

Telephone (_____) _____

Delivery is not made until payment is received

To order by phone call **1 800 633-4931**

or (205) 995-1567 outside the US and Canada

❷ ❑ My payment for $_____ is enclosed.

OR

❑ Bill me. My library's purchase order number is _____ .

❸ Send this order and payment to:

INTERPERIODICA

P.O. Box 1831

Birmingham, AL 35201-1831

Or fax this order form to (205) 995-1588

Order now for immediate delivery. All journals will be published in 1993!

Russia's leading geological research brought to you in English direct from Moscow by INTERPERIODICA

PETROLOGY

Editor-in-Chief
Vilen A. Zharikov
Director of the Institute of Experimental Mineralogy, Russian Academy of Sciences

PETROLOGY is a journal of magmatic, metamorphic, and experimental petrology, mineralogy, and isotopic geochemistry. The journal offers comprehensive information on all multidisciplinary aspects of theoretical and experimental petrology. By giving special consideration to studies on the petrography of different regions of the former Soviet Union, PETROLOGY will provide all readers with a unique opportunity to refine their understanding of the geology of the vast territory of the Eurasian continent.

Coverage includes:

- petrology of magmatic, metamorphic, and metasomatic rocks in continental and oceanic environments
- experimental petrology and mineralogy
- geothermobarometry
- physiochemical analysis of mineral assemblages
- isotopic geochemistry and geochronology.

Bimonthly. 1993 Vol. 1. ISSN 0869-5911.
North America, $540 Rest of World, $576

STRATIGRAPHY AND GEOLOGICAL CORRELATION

Editor-in-Chief
Boris S. Sokolov
Academician and Adviser of the Presidium, Russian Academy of Sciences

Previously combined with PETROLOGY in the original Russian, this new journal fully explores stratigraphy and the chronological features of the world's geological record. STRATIGRAPHY AND GEOLOGICAL CORRELATION is the only journal in the world to provide comprehensive coverage of fundamental and applied aspects of stratigraphy and the correlation of geological events and processes in time and space. Articles are based on the results of multi-disciplinary research and cover a wide area of interests including:

- problems of general and regional stratigraphy of continents and the oceans sedimentary cover
- theory and methods of stratigraphic analysis
- geochronology including isotopic geochronology
- the evolution of the biosphere
- development of sedimentary basins
- aspects of geological correlation and global secular changes in ecosystems.

Bimonthly. 1993 Vol. 1. ISSN 0869-5938.
North America, $540 Rest of World, $576

GEOLOGY OF ORE DEPOSITS

Editor-in-Chief
Nikolai P. Lowerov
Director of the Institute of Geology of Ore Deposits, Petrology, Mineralogy and Geochemistry, Russian Academy of Sciences

Founded in 1959, GEOLOGY OF ORE DEPOSITS is one of only a few devoted to the study of metallic and non-metallic minerals and the conditions of their formation and distribution. Since restrictions on publishing classified geological data have been lifted, more complete and concrete findings on important deposits throughout the former USSR — including such unique ore districts as Noril'sk, Udokan, and Muruntau — are now available. Also included are articles by foreign contributors on strategic deposits through the world. GEOLOGY OF ORE DEPOSITS covers:

- economic geology of metal and non-metal ore deposits
- geological environments of ore formation
- mineral composition of ores
- experimental studies of ore-forming systems
- ore-forming processes and their evolution
- geochronology of ore-forming processes and their evolution
- criteria of forecasting and estimating mineral deposits.

Bimonthly. 1993 Vol. 35. ISSN 0016-7770.
North America, $540 Rest of World, $576

INTERPERIODICA

pledges to give you the best quality and service:

- Immediate access to unique scientific research
- Superior translations — produced by a carefully chosen team working closely, in Moscow, with the Russian authors and editors
- High quality printing in North America...journals attractively designed and printed with high standards to ensure rapid delivery
- Centralized customer service — responsive attention to your needs through our North American subscription office
- Reasonable prices...and now offering special discounts
- Full no-risk, money-back guarantee if you are not completely satisfied

▼

ALL 1993
JOURNALS WILL BE
PUBLISHED IN 1993!

▲

For more information call
1 800 633-4931 (US and Canada)
or 205 995- 1567 (Rest of World)
or write
INTERPERIODICA
P.O. Box 1831
Birmingham, AL 35201-1831